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SEDIMENT TRANSFER FROM GRAVEL- BED RIVERS TO BEACHES

thesis

submitted to the faculty of science

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by

SALLY L.G. GEMMELL

APRIL, 2000

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ABSTRACT

Sediment transfer from gravel-bed rivers to beaches

Sally L.G. Gemmell

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A morphological approach was used to quantify the transfer of gravel-sized ($> 2\text{mm}$) sediment from one river to the coast and its subsequent redistribution within the coastal zone. The study investigates sediment transport in wandering gravel-bed rivers, river mouth processes and sediment transport on gravel beaches. The research was largely field based and used results from repeat morphologic surveys of the lower, wandering gravel-bed reach of the River Spey, the Speymouth delta and gravel beaches of Spey Bay, north-east Scotland.

The supply of gravel to the river mouth was episodic, with transport rates varying from $41\,000 \pm 6\,000 \text{ m}^3 \text{ a}^{-1}$ (1993-1994) to $6\,000 \pm 4\,000 \text{ m}^3 \text{ a}^{-1}$ (1997-1999). Transport rates were not directly related to flow conditions because the availability of sediment for transport was critical. Sediment was mobilised according to the recent local history of erosion, deposition and channel adjustment and not only the magnitude of the flood.

Delivery to the coast of this episodic sediment supply was made more variable depending on the operation of the delta. A gravel spit complex extended westwards across the river mouth at a mean rate of 150 m a^{-1} between 1997 and 1999. This resulted in temporary storage of sediment in the extending spit which had implications for the downdrift coast. Cycles of accretion and erosion were created in the lee of the spit which were most significant at the river mouth before being propagated downdrift.

Volumetric information obtained from successive beach profile surveys indicated that zones of accretion and erosion were spatially and temporally variable along the 16km coastline of Spey Bay. This variability was caused by the passage of pulses (or slugs) of sediment which moved alongshore in response to variations in sediment supply (e.g. episodic delivery of fluvial sediment, river mouth processes, beach feeding and storms). It is argued that gravel sediment moves alongshore as slugs by a given distance depending on the magnitude of storm events. Travel distances of ca. 2-3km were recorded following a major storm event in March 1998. This has important implications for the management of gravel beaches, as erosional zones (or "problems") are not static and vary in space and time.

Fluvial, deltaic and coastal volume changes were combined to estimate a short-term (3 year) sediment budget for the system. A medium-term budget (100 year) was also compiled from map analysis and river-modelling studies. Both budgets showed a net loss of sediment from the system, indicating a system that has been erosional for at least the last century.

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1. INTRODUCTION

Although little quantitative information is available about the supply of gravel-sized sediment ($> 2\text{mm}$) from rivers to coasts, its importance is recognised and is crucial for understanding the long-term stability of gravel coasts (e.g. Forbes et al. 1995). In particular, with climate change scenarios forecasting changes in sea-level (Shennan 1993) and in the frequencies of both river flooding and storm incidence (e.g. Beven 1993; Smith and Bennett 1994), the dynamics of river-coast exchanges require urgent investigation.

Exchanges of gravel-sized sediment at the fluvial-coastal interface remain relatively understudied despite recent research in New Zealand (Kirk 1991; Shulmeister and Kirk 1997), whereas sand-sized sediment exchanges have received far greater attention (e.g. Hicks and Inman 1987; Jimenez and Sanchez Arcilla 1993; Hicks et al. 1999). There is evidence that river mouth processes can cause large-scale irregular variations in the regional longshore transport regime, leading to temporal and spatial variations in the rate of coastal erosion along the downdrift coast (Hicks and Inman 1987; Kirk 1991) although the reasons for this are not yet fully investigated. River mouth processes result from the complex interaction of fluvial and coastal processes operating at different timescales and so the frequency and magnitude of both river flooding and coastal storms are likely to be important factors governing sediment exchanges.

On the other hand, the dynamics of sediment transfers and storage in gravel-bed rivers are relatively well understood (e.g. Church and Jones 1982; Lane et al. 1995; Madej and Ozaki 1996; Wathen and Hoey 1998). However, it has recently been observed that the basic concept of bedload transport in the deep channel bed may be conceptually inappropriate in large wandering channels (McLean and Church 1999). In large wandering gravel-bed rivers mobile sediment is derived from distinct points of erosion along the channel and moves to separate points of deposition (Carson and Griffiths 1989; McLean and Church 1999). This relationship between sediment transport and river morphology has been used to develop a sediment budget approach to quantify sediment transfers on such rivers (Ashmore and Church 1999) which will be further advanced in this study.

The coastal literature abounds with studies concerning sand transport (e.g. Komar 1990; Komar 1996) while studies of gravel transport are relatively scarce. Recent advances in the use of gravel tracers has allowed longshore gravel rates to be quantified and inferences made about the nature of transport (e.g. Matthews 1980; Bray et al. 1996). Evidence suggests that gravel moves along the beach in slugs (Single and Hemmingsen 2000), and

not as individual grains as on sand beaches, although further work is required to verify this model.

Arising from the above, this study addresses three main gaps in the current geomorphological literature:

- sediment transfer volumes (of gravel-sized sediment) from rivers to the coast;
- the dynamics of sediment transfers and storage at the fluvial-coastal interface and;
- the dynamics of gravel transport and storage in the coastal zone.

Specific objectives are to:

1. Quantify fluvial sediment transport rates and sediment storage in the lower reaches of a wandering gravel-bed river;
2. Quantify sediment transport, exchanges and storage at a high energy fluvial-coastal interface and assess the relative importance of fluvial and coastal processes in sediment delivery;
3. Quantify coastal sediment transport rates and sediment storage on high energy gravel beaches;
4. Construct a contemporary sediment budget for a large fluvial-coastal gravel system using results from objectives 1, 2 and 3;
5. Construct a sediment budget for the same system over 100 year and 10 000 year timescales and assess the effect of changing sediment supply and delivery on system evolution;
6. Examine the relationship between fluvial-coastal sediment exchanges with river flows (i.e. floods) and wave conditions (i.e. storms) and determine other morphological, structural and sedimentological factors that may be influence transfer rates; and
7. Develop a conceptual model to account for sediment transfers from wandering gravel-bed rivers to the coast.

This research requires the identification of a dynamic site which permits the natural transfer of gravel-sized sediment from the fluvial system to the coast. Ideally, the site should be subject to minimal human modification and consist of an active and dynamic gravel-bed river discharging into an active gravel coastline. The lower River Spey as it enters the Moray Firth at Spey Bay, north-east Scotland fits these requirements. The dynamics of the gravel-bed river result in a constantly changing channel and gravel supply

(Lewin and Weir 1977; Riddell and Fuller 1995) which reaches the coast via a dynamic spit/mouth complex (Grove 1955). Wave processes redistribute the supplied gravel to the remainder of the beach making this an ideal site to study gravel transport.

One of the most useful concepts in coastal research and management is the “budget of sediments” (Komar 1996). This is in essence an application of the principle of conservation of mass to littoral sediments (Bowen and Inman 1966) and can be used to quantify littoral transport rates (e.g. Drapeau and Mercier 1987; Deruig and Louisse 1991). The sediment budget approach has also been used to quantify fluvial sediment transfers over a wide range of spatial and temporal scales (e.g. Ferguson and Ashworth 1992; Martin and Church 1995). This study applies a sediment budget approach to quantify gravel transfers from rivers to the coast and so integrate the two systems.

Structure of the thesis

Chapter 2 reviews the current research at the fluvial-coastal interface, with particular focus on gravel systems. The relevant literature concerning fluvial bedload transport and coastal sediment transport is reviewed and gaps identified. The sediment budget approach is evaluated and its applicability to quantifying sediment transfers from rivers to beaches assessed. Chapter 3 introduces the field site and describes the geomorphological, hydraulic and sedimentological characteristics of the river and coast. The methods of field data collection and analysis are outlined in Chapter 4 and a detailed error analysis of the temporal and spatial density of field survey required to accurately quantify sediment transfers is presented. Chapters 5 and 6 present the results and analysis of this research. Sediment storage and transfers in the fluvial and deltaic system are quantified in Chapter 5, while the data pertaining to coastal sediment transfers are presented and discussed in Chapter 6. Chapter 7 summarises the results of the main analytical chapters and amalgamates the data to produce short-term (3 year), medium-term (100 year) and long-term (10 000 year) sediment budgets for the Spey River and Bay system. The nature of sediment exchanges in the fluvial, deltaic and coastal systems are examined and a conceptual model describing the transfer of gravel-sized sediment from rivers to the coast is proposed. Conclusions are summarised in Chapter 8, where the wider implications of this research are evaluated and areas for further research identified.

2. FLUVIAL AND COASTAL SEDIMENT TRANSFERS

Many researchers have utilised sediment budget approaches to quantify sediment transfers within fluvial and coastal environments over a wide range of temporal and spatial scales (e.g. Clayton 1980; Dietrich et al. 1982). This chapter introduces this approach and assesses its suitability for the quantification of gravel transfers from rivers to beaches. A review of previous sediment budget studies is presented and their applicability at a variety of scales and in different environments is evaluated.

The dynamics of sediment transfers and storage in gravel-bed rivers are relatively well understood (e.g. Church and Jones 1982; Lane et al. 1995; Madej and Ozaki 1996; Wathen and Hoey 1998). However, the prediction of bedload transport rates using traditional empirical formulae is problematic (Davies 1987; Gomez and Church 1989) and a sediment budget approach to quantifying transport rates in gravel-bed rivers has recently been advanced (e.g. Ashmore and Church 1999). In contrast, the dynamics of sediment transfers and storage on gravel beaches are less well understood and the prediction of sediment transport on gravel beaches is equally problematic (Dolan et al 1987; Carter and Orford 1993). Indeed, it is debatable whether any of the standard beach transport formulae (e.g. the CERC 1984 formula for longshore sand transport), all of which were developed for sand sized material, can be extended over the several orders of magnitude needed to accommodate coarse clastic material (Carter and Orford 1993). As a result, coastal researchers have applied the sediment budget approach to estimate coastal sediment transport at a variety of scales in a range of environments (e.g. Comber 1993; Bray et al. 1995; Komar 1996). Sediment transfers at the fluvial-coastal interface, particularly of gravel sized sediment, are very poorly understood. This chapter critically reviews the relevant literature.

2.1 Conceptual framework - the sediment budget approach

A sediment budget is defined as the quantification of various types of movement and storage of sediment within a landscape unit (Rawat 1987) and has been used over a wide range of temporal and spatial scales in the fields of both fluvial (Section 2.1.1) and coastal geomorphology (Section 2.1.2).

2.1.1 Fluvial sediment budgets

The development of a within channel sediment budget for a gravel-bed river utilises principles derived from basin scale approaches (e.g. Dietrich et al. 1982) to describe and predict sediment transfers. A sediment budget for a drainage basin provides a quantitative accounting of the rates of production, transport, storage and discharge of sediment. Dietrich et al. (1982) describe three main requirements for the construction of a sediment budget: (1) recognition and quantification of transport processes; (2) recognition and quantification of storage elements; and (3) identification of linkages between transport processes and storage sites.

2.1.1.1 Reservoir theory

Sediment budgeting involves the calculation of changes in volume (or mass) of stored sediment from one reach to the next. However, even with extensive field measurements, it is difficult to assess the net sediment input and discharge out of a reach during a defined time period (Kelsey et al. 1987). This problem is complex because bedload size sediment in the channel and on adjacent depositional surfaces is not all equally accessible to transport. For example, some sediment within a reach is readily available for transport in the active channel and, at the other extreme, sediment in vegetated terraces is only transported by infrequent, high magnitude floods. The classification of sediment into reservoirs or stores (of activity) within a reach allows the accessibility of sediment to transport to be identified.

Division of the reach into discrete storage units and the quantification of storage elements is essential to the calculation of a sediment budget (Eriksson 1971; Bolin and Rodhe 1973; Dietrich et al. 1982). A store is defined by Wathen (1995) as “a volume or area of sediment bordered by numerically defined boundaries representing a specific range of potential transfer conditions” (p107). Reservoir theory (Eriksson 1971, Nakamura 1986) provides a framework for characterisation of these stores according to sediment age (and hence activity) using descriptive variables such as residence time to provide a quantitative assessment of storage within the reach.

A number of variables have been used to classify within channel sediment storage including elevation, potential activity, vegetation age and the distribution of sediment transfers (e.g. Kelsey et al. 1987; Madej 1987; Nakamura et al. 1995; Wathen et al. 1997). Kelsey et al. (1987) mapped and classified all sediment stored in the main channel of Redwood Creek into one of four reservoirs: active, semi-active, inactive, and stable (Figure

2.1) based on the relative position and elevation of stored sediment in comparison to the active channel, the density and age of vegetation growing on the deposit and the analysis of sequential aerial photographs.

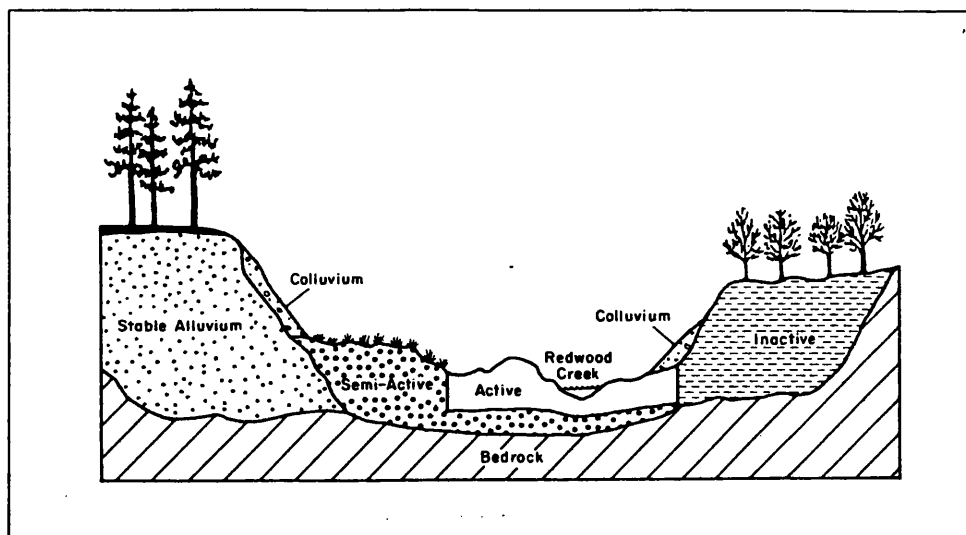


Figure 2.1: Classification of the four sediment reservoirs based on the relative mobility of sediment (source: Kelsey et al. 1987).

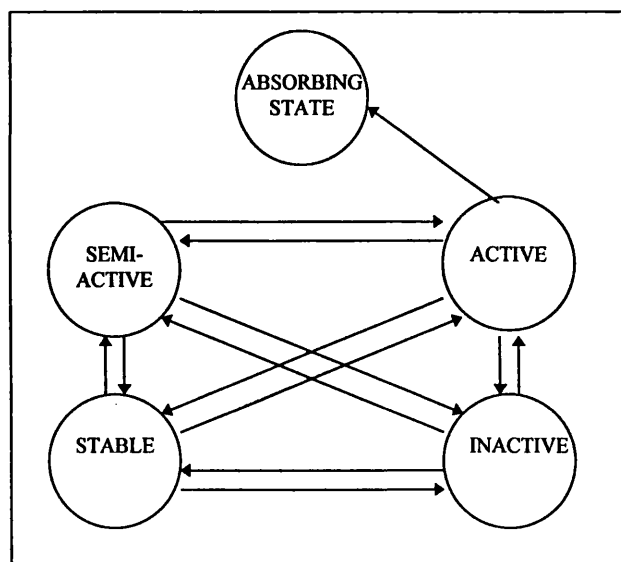


Figure 2.2: State diagram showing the possible interchanges between the four sediment reservoirs. Note that sediment can only exit (or be absorbed) from the system through the active reservoir (source: after Kelsey et al. 1987).

The transit times and fate of sediment entering a channel system are sensitive to exchanges between reservoirs, each having different residence (or turnover) times (Kelsey et al. 1987; Madej 1987). The state diagram (Figure 2.2) indicates the possible transitions between the four sediment reservoirs in a given reach. Sediment may also exit (or be absorbed) from the system to a downstream reach or to the coast although this can only occur via the active

reservoir (Kelsey et al. 1987). Hoey (1996) notes that there are two ways in which sediment can ‘move’ between reservoirs. *Dynamic transfers* involve the physical movement of sediment from one reservoir to another. *Static transfers* occur when a change in the water surface elevation relative to the sediment necessitates reclassification to a different reservoir. This occurs frequently and at different rates in active gravel-bed rivers. For example, as a channel migrates across its floodplain, some sediment gradually becomes further removed from the channel and thus becomes part of a less active reservoir without moving. Rapid static transfers can occur during channel avulsions.

2.1.1.2 *The morphological approach*

The morphological approach to quantifying sediment transfers is reviewed thoroughly by Ashmore and Church (1999). Gravel transport rates can be estimated from volumetric morphological data using two main approaches (Goff and Ashmore 1994): (i) the step length approach (Neill, 1987; Carson and Griffiths 1989; Ferguson and Ashworth 1992); and (ii) the sediment budget approach (Griffiths 1979; Ferguson and Ashworth 1992; Martin and Church 1995).

(i) *Step length approach*

The step length approach was first developed by Neill (1971, 1987) for single-thread, meandering channels based on the measurement of bank retreat rates combined with the assumption that the eroded sediment travels a known distance to the deposition site (Figure 2.3). The volumetric transport rate is (from Neill 1987):

$$Q_s = L \cdot h \cdot \frac{dE}{dt} \quad (2.1)$$

where L is the average length of travel between erosion and deposition (i.e. the step length), h is the bank height and dE/dt is the bank recession rate. For meandering channels the step length was taken to be one half the meander wavelength (Neill 1987) although it was recognised that this may need to be redefined for different channel morphologies.

Neill’s (1971, 1987) method has been adapted for wandering gravel-bed rivers (e.g. Ferguson and Ashworth 1992; McLean and Church 1999). Volumes of upstream erosion are matched with similar volumes of deposition downstream and the distance between the centroids of each volume represents the transfer distance (‘step length’). Combining this with a known or assumed time period over which the transfers occur, yields a volumetric transport rate.

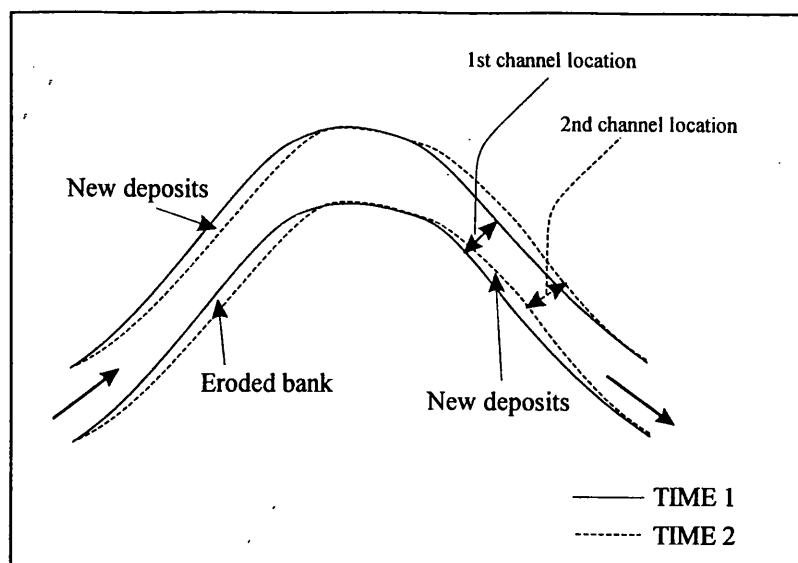


Figure 2.3: The step length approach to quantifying sediment transfers (source: after Neill 1987)

Carson and Griffiths (1989) applied this method to the braided Waimakariri River, New Zealand. Superimposition of two consecutive photos indicated areas of the floodplain which were transformed (a) from deep water to shallow water; (b) from deep water to dry river bed; (c) from shallow water to deep water; (d) from dry, unvegetated river bed to deep water; and (e) from vegetated islands of floodplain to deep water in the given time period. This method thus accounts for both static and dynamic sediment transfers (cf. Hoey 1996). The areas of scour or deposition in each category were digitised and the volumes determined using a value for the depth of scour or fill estimated from a standard survey of the river reach. Volumes of erosion were combined with the average step length to estimate transport rates, which were corroborated with rates calculated using equilibrium transport functions (Carson and Griffiths 1989).

(ii) Sediment budget approach- the concept of continuity

Reach sediment storage change estimates can be used to infer bedload transport rates using the concept of sediment continuity. This approach is applicable whether or not clear erosion/deposition cells can be identified (Ferguson and Ashworth 1992). The construction of a sediment budget involves the quantification of sediment inputs, outputs and storage changes in a defined reach, based most simply on the equation of sediment continuity:

$$S_o = S_i - \delta S \quad (2.2)$$

where S_i is bed material input, δS is change in storage and S_o is bed material output. If two of these terms are known then the third can be calculated within the margin of error of the known terms. If each of these variables is measured over a finite time then the equation becomes (from Martin and Church 1995):

$$Q_o = Q_i - (1 - \rho) \delta S / \delta t \quad (2.3)$$

where Q_i and Q_o are the volumetric transport rates into and out of, respectively, a reach per unit time (δt) and ρ is the porosity of the deposited sediment. The time unit is the time between surveys or, in some cases, the time when the flow was competent to move sediment. In practise, Q_i and Q_o are often large numbers, for example Q_i is $156\,800\text{ m}^3\text{a}^{-1}$ at Vedder Crossing, Canada (Martin and Church 1995), so that acceptable percentage errors for these terms may permit relatively large percentage errors in δS . This is explored further in Section 4.6.

An independent estimate of Q_i or Q_o must be obtained at one section in order to determine its value at all other sections, as quantity Q_i into one reach is Q_o from the next upstream reach (Ferguson and Ashworth 1992; Martin and Church 1995). Independent predictions of bedload transport rates using conventional formulae (such as Einstein, Parker and Ackers-White-Day) are problematic and give inconsistent results (Gomez and Church 1989). In addition, direct field measurements of bedload transport in large gravel-bed rivers are often logistically impractical. Errors in the estimate of Q_i or Q_o must be quantified, as these errors will propagate either up or downstream. If no direct measurements of bedload transport are available a lower bound can be set by the requirement of non-negative transport at all sections (e.g. Griffiths 1979) or a downstream boundary condition of zero gravel transport can be assumed, as in the budget calculated for the lowermost gravel-bed reach of the Fraser River, Canada (McLean 1990). The distinct gravel-sand transition on the Fraser (McLean 1990) verifies the zero transport assumption in this case. For other rivers the budget can be closed using any independent estimate of Q_i or Q_o . Even if the closing estimate is wrong, the relative transport rates along the river remain the same (Martin 1992) and errors may be small compared to errors in the storage change estimates.

Estimation of Reach Storage Change

Change in sediment storage (δS) for a given reach can be calculated either from repeat surveys of the same cross-sections (Ferguson et al. 1992; Ferguson and Ashworth 1992; Martin and Church 1995; Madej and Ozaki 1996), from digital elevation models of successive reach surveys (Lane et al. 1994) or from repeat planimetric mapping of channel change (Ham 1996; McLean and Church 1999).

(a) Cross-section surveys

Successive cross-section surveys are overlaid to obtain the net change in cross-section area as (from Martin and Church 1995):

$$\delta A = \sum_{i=1}^{n-1} \delta A_{(i,i+1)} = \sum_{i=1}^{n-1} \frac{\delta z_i + \delta z_{(i+1)}}{2} d_{(i,i+1)} \quad (2.4)$$

where $i, i+1$ represent two successive points on the survey line, δz is the change in elevation between surveys and d is the distance between the two points. Net volumetric changes are then calculated between sections on the assumption that the change in area at a cross-section is representative over the distance between it and the half distance to each adjacent cross-section:

$$\delta V = \frac{\delta A_j + \delta A_{(j+1)}}{2} L_{(j,j+1)} \quad (2.5)$$

where δA_j is the change in area at cross-section j , δA_{j+1} is the change in area at the next upstream cross-section, and $L_{(j,j+1)}$ is the distance between the two cross-sections. Once the volumetric change for each survey unit is determined and if the bedload transport rate is known at one place along the channel, calculations can be extended upstream or downstream using the sediment budget approach (equation 2.3).

(b) Reach Scale Surveys

Surveys of the detailed reach morphology can be obtained by conventional survey (e.g. Wathen and Hoey 1998), photogrammetrically (e.g. Lane et al. 1995; Heritage et al. 1998), or a combination of both (e.g. Lane et al. 1994). The individual data points from reach scale surveys are generally input into GIS software to create a digital elevation model (DEM) and the *change* in sediment storage for a given reach and time step is calculated as the difference between successive DEM surfaces (e.g. Lane et al. 1994; Wathen and Hoey 1998; McLean and Church 1999).

Reach scale surveys have advantages over fixed cross-section surveys (Figure 2.4). Successive surveys do not reoccupy the same points and so survey density can be re-adjusted to the changing terrain and evolve along with the landform (Lane et al. 1994; Ashmore and Church 1999). In addition, cross-section surveys are criticised as they require a trade off between time spent collecting data at higher densities and over wider areas and the frequency of return to the same points to measure how the landform is changing (Lane et al. 1995).

The problem of sediment throughput, where sediment moves through the reach with little or no morphological signature (Carson and Griffiths 1989; Lane and Richards 1997) is inherent in the morphological method of bedload transport estimation. There are two types of throughput: (a) where the sediment goes directly through the reach; and (b) where

sediment coming into a reach replaces sediment leaving. The first problem can be solved by increasing the reach length. There is no simple solution to the second and estimates of transport rates using this method should always be considered lower-bound (Lane et al. 1994). The potential effect of this problem is minimised using reach scale surveys compared to fixed cross-section surveys (Figure 2.4) as this will avoid sediment transfers from one ‘storage element’ bar (Church and Jones 1982) to the next being undetected.

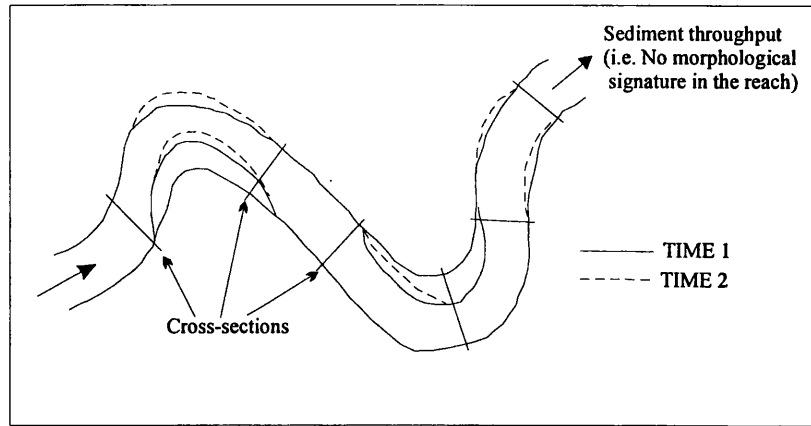


Figure 2.4: Fixed cross-section surveys compared to reach scale surveys for sediment budgeting. Cross-sections at the locations shown will “miss” all the morphological changes that occurred between time 1 and 2. Surveys of the entire reach morphology will capture these changes.

(c) Planimetric mapping of channel changes

Ham (1996) estimated reach storage changes and bedload transport rates on the Chilliwack River, British Columbia using a planimetric mapping technique. Areas of erosion and deposition were identified by overlaying successive aerial photographs of the channel. These were converted to volumes by multiplying by the depth of the mobile sediment (or scour depth) (Ham 1996). This is the maximum vertical mobilisation of the bed material as determined from scour chains or repeat cross-section surveys (Madej 1995). Accurate definition of this depth is required for successful application of the planimetric approach.

Planimetric methods do not account for changes in the channel bed elevation. This amounts to assuming that the channel thalweg does not participate significantly in sediment exchanges but simply is sequentially buried and exhumed as bar growth and bank erosion occur (McLean and Church 1999). The concordance of gravel transport estimates using this approach with direct measurements of gravel transport on the Fraser River (McLean and Church 1999) provides evidence that this method is not significantly negatively biased and for the Fraser River, at least, the assumption that the channel thalweg does not participate in sediment exchanges appears to be sustained. Therefore sediment

throughput which has no morphological signature may be minimal (relative to the actual transfers) on large wandering gravel-bed rivers.

Variations of this method have been successfully applied to several large, wandering and braided gravel-bed rivers (e.g. the Waimakariri, New Zealand (Carson and Griffiths 1989), the Fraser, Canada (McLean 1990; McLean and Church 1999) and the Chilliwack, Canada (Ham 1996)) all of which are similar in scale and characteristics to the lower River Spey.

Spatial and Temporal Survey Density Issues

Successful application of the sediment budget approach using equation 2.3 requires accurate estimation of the change in reach storage (ΔS) through time. The uncertainty of the storage change estimates should always be assessed whether cross-section surveys, reach scales surveys or planimetric techniques are used. The spatial and temporal density of re-survey or re-mapping is important.

Cross-section spacing should be determined on the basis of study aims and system scale. Cross-sections have been used to estimate reach storage changes at many scales: from small and complex proglacial streams (e.g. Ferguson and Ashworth 1992; Goff and Ashmore 1994; Lane et al. 1994) up to large, wide braided rivers (Griffiths 1979). In large channels, in which downstream patterns of erosion and deposition occur at large scales, large cross-section spacing may be appropriate (e.g. Griffiths 1979; McLean 1990), while in the complex proglacial channels cross-section spacing must be small (Lane et al 1994). To date there has been no systematic analysis of the error associated with variation in cross-section spacing (Ashmore and Church 1999).

Cross-section spacing varies from 1m (Lane et al. 1994; Wathen 1995) up to 2km (Griffiths 1979; McLean 1990) depending on system scale. In dimensionless form (= spacing/mean channel width), spacing varies between 0.05 (Lane et al. 1994) and 9 times the mean channel width (Madej and Ozaki 1996), although typically cross-section spacing is ca. 2-3 (e.g. Griffiths 1979; Ferguson and Ashworth 1992; Martin 1992; Martin and Church 1995; McLean and Church 1999).

The temporal density of survey (Δt in equation 2.3) also affects the transport rate obtained from morphological calculations. Transport rates estimated using this method are inherently negatively-biased and represent lower-bound estimates due to compensating erosion and deposition between surveys (Goff and Ashmore 1994; Lane et al. 1994; Martin and Church 1995). For example, any record of sediment stored and then re-entrained within a period shorter than the time resolution of the study is lost. This limits the

usefulness of this approach to rivers which experience compensating scour and fill within a relatively short term. However, the close consistency between gravel transport rates computed using morphological methods with direct measurements on the lower Fraser River (McLean and Church 1999) indicate that no significant negative bias is introduced using morphological methods at a time step of 32 years. Therefore the temporal density of re-survey is a function of river size, as morphological adjustments take longer to achieve in larger rivers. Observations on large rivers (Carson and Griffiths 1989; McLean and Church 1999) suggest that mobile sediment is derived from distinct points of erosion along the channel and moves to separate points of deposition. The resulting channel realignment causes relocation of localised erosion and deposition zones over extended periods of time, of the order of 10 years in the Fraser (McLean and Church 1999).

2.1.2 Coastal sediment budgets

Since the formulation and initial application of the sediment budget concept in the coastal zone by Bowen and Inman (1966) analysis of sediment budgets have been undertaken in a variety of coastal settings, albeit mainly for sandy beaches (Komar 1996). The development of a coastal sediment budget involves quantifying the gains and losses of sediment within a coastal cell (Bowen and Inman 1966). This can then be compared with the observed rate of beach changes reflected in profile erosion or accretion (Komar 1996). A sediment budget approach requires estimation of all inputs to and outputs from the coastal system, including the transfer of sediment from rivers to beaches (Table 2.1).

Credit	Debit	Balance
Longshore transport into cell	Longshore transport out of cell	Beach accretion or erosion
River transport	Wind transport away from the beach	
Cliff erosion	Offshore transport	
Onshore transport	Deposition in submarine canyons	
Wind transport onto beach	Solution and abrasion	
Beach nourishment	Beach Mining	
Biogenic deposition		

Table 2.1: The Budget of Littoral Sediments (source: Bowen and Inman 1966)

The primary units of sediment budgets are “littoral drift cells”, which are self contained entities within which sediment circulates (Bray et al. 1995; Komar 1996). Coastal areas can be divided into a series of littoral cells according to morphological and process information. The simplest littoral cell is a pocket beach, isolated by rocky headlands. It is often useful to consider the coast as a hierarchy of cells, where smaller sub-cells are analysed within the large-scale littoral cell (Komar 1996). For example, Bray et al. (1995)

defined a series of littoral cells and sub-cells for the south coast of England drawing on diverse sources of information documenting the movement of gravel along the beaches and in the offshore.

The concept of sediment continuity

Coastal sediment budgets and longshore transport rates can be estimated using the concept of sediment continuity. This approach has been used mainly in computer models of shoreline prediction (e.g. Uda and Saito 1987; Hanson and Larson 1987; Komar 1998) where the shoreline is split into a series of cells (Figure 2.5a) and the end of each cell terminates in a schematic beach profile (Figure 2.5b). The basis of any shoreline model is the evaluation of the quantities of sediment entering and leaving the cell, and the resulting changes in the shoreline position (or beach volume) due to the balance of input and output (Komar 1998). Littoral drift is usually the main cause of sediment moving from one cell to another. Applying the concept of sediment continuity, the net volume of sediment gained or lost from cell i over time t is (from Komar 1998):

$$\Delta V_i = (Q_i - Q_o \pm Q_r) \Delta t \quad (2.6)$$

where Q_i and Q_o are the littoral transport rates into and out of cell i , respectively and Q_r is a term accounting for the various other sources or sinks of beach sediment.

For shoreline prediction it is desirable to express ΔV_i as an actual change in shoreline position, that is, as a change in the length y_i of the cell (Figure 2.5a). From the geometry of the cell shown in Figure 2.5b:

$$\Delta V_i = d \Delta y_i \Delta x \quad (2.7)$$

The height d (Figure 2.5b) is chosen to yield the correct correspondence between ΔV_i and Δy_i and depends on the nature of the beach profile (Komar 1998). Combining equations 2.6 and 2.7 gives:

$$\Delta y_i = (Q_i - Q_o \pm Q_r) \frac{\Delta t}{d \Delta x} \quad (2.8)$$

for a change in shoreline position of cell i as a function of sediment inputs and outputs. Decreasing the finite terms to their limits yields (Komar 1998):

$$\frac{dy}{dt} = -\frac{1}{d} \frac{dQ_s}{dx} \quad (2.9)$$

where Q_s is the longshore sediment transport rate. This relationship highlights the dependence of the time-rate of shoreline change, dy/dt , on the longshore gradient of the

littoral drift, dQ_s/dx . For example if Q_s is increasing in the longshore direction (i.e. $dQ_s/dx > 0$), the model predicts a shoreline retreat (i.e. $dy/dt < 0$) as more sediment leaves the cell than enters it. A decreasing Q_s results in a shoreline advance.

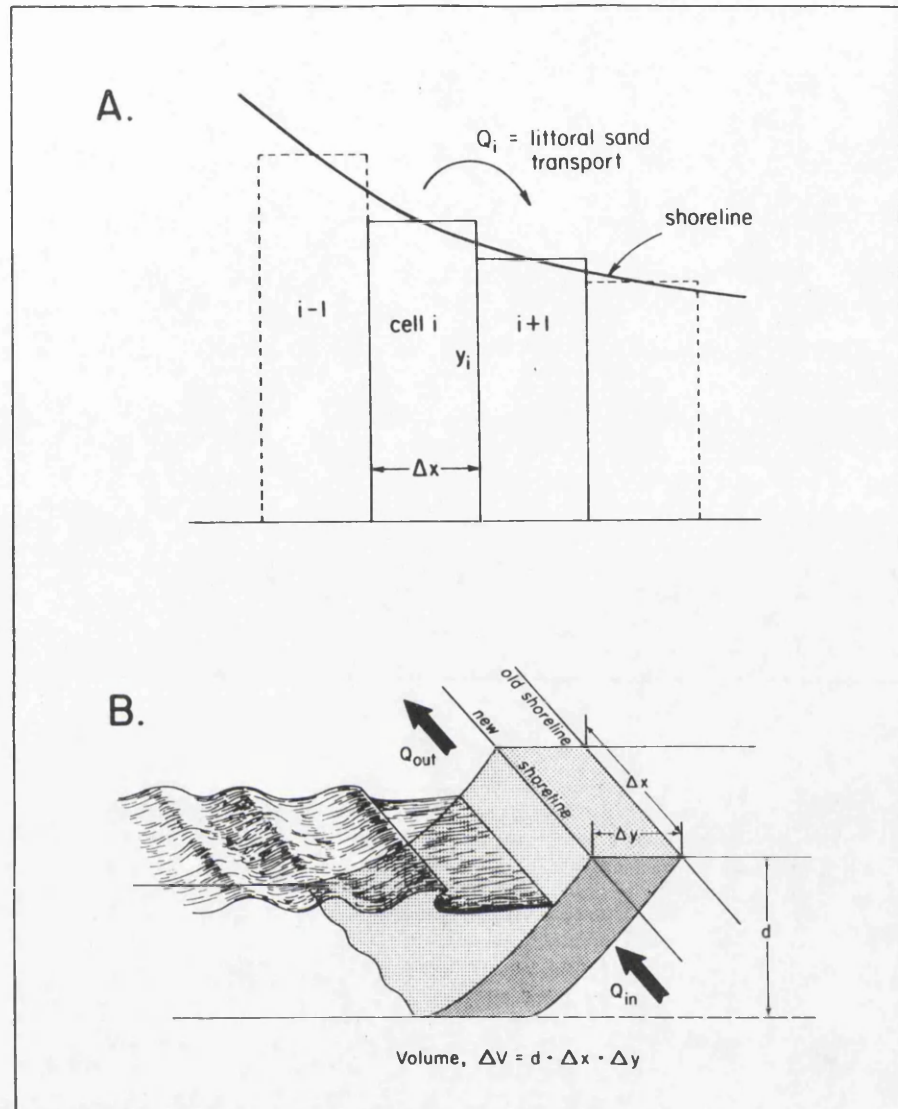


Figure 2.5: (a) The shoreline is divided up into a series of cells of width Δx and variable y_i lengths beyond some baseline. (b) One shoreline cell demonstrating how a change in sand volume ΔV is produced by the littoral drift into and out of the cell and how this results in a shoreline change, Δy (source: Komar 1973, 1983).

Many numerical models of shoreline change are based on this concept and use the variation in potential longshore transport rate to predict shoreline changes (e.g. Greenwood and McGillivray 1978; Davidson-Arnott and Pollard 1980; Komar 1983; Komar 1998). However, few studies have used the concept of sediment continuity the other way around (i.e. using *actual* changes in beach cell volumes to estimate longshore transport rates). To do this equation 2.6 can be re-written as:

$$Q_o = Q_i - (1 - \rho) \delta S / \delta t \pm Q_r \quad (2.10)$$

where ΔS is the change in sediment storage in a given beach cell over a given time (Δt) and ρ is the porosity of the sediment. This is essentially the same as equation 2.3 for fluvial sediment transport rates. To quantify longshore transport rates using this approach requires the accurate estimation of beach cell storage changes, an independent prediction (or assumption) of Q_i or Q_o at one cell boundary and knowledge of the direction of littoral drift.

Estimation of beach cell storage changes

(1) Analysis of shoreline change

Quantification of the actual gain or loss of sediment at the coast is often based on the analysis of sequential aerial photographs and maps (e.g. Allen 1981; Leatherman et al. 1987). From this analysis the area of beach accretion or recession can be quantified and, in some cases, converted into a beach volume.

(2) Equilibrium beach profiles

More directly, the equilibrium beach profile concept (Bruun 1962) can be used to estimate the change in beach volume from a given shoreline change. The ‘Bruun Rule’ considers that as sea-level (or water level) rises an equilibrium profile is maintained as the shoreline is displaced landward and upward (Figure 2.6). It is assumed that the volume of sediment eroded from the subaerial part of the profile will equal the volume deposited on the lower shoreface.

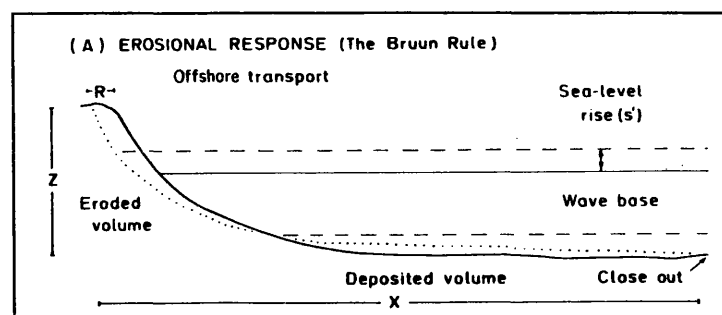


Figure 2.6: The ‘Bruun Rule’ (Bruun 1962) (source: Carter 1988). The basic ‘Bruun Rule’ is given by the equation: $R = \frac{xs'}{z}$ where R is the shoreline recession, x is the profile width, s' is the sea-level rise and z is the profile depth.

The volume change in the lower shoreface is assumed to take place down to a specified depth (Figure 2.6). This is often called the “depth of effective motion” (Dean and Maurmeyer 1983) or the “depth of closure” (Hanson and Larson 1987) and represents the maximum depth at which sediment is moved. Adequate definition of the closure depth is a

source of significant uncertainty when using the ‘Bruun Rule’ (Carter 1988; Mimura and Nobuoka 1995).

The one-line theory of shoreline change is based on the equilibrium profile concept and assumes that erosion or accretion of a beach results in a pure translation of the beach profile and the bottom profile moves in parallel to itself without changing shape (Hanson and Larson 1987). It is also assumed that longshore transport takes place uniformly over the beach profile down to a limiting depth (the depth of closure) (Hanson and Larson 1987). If short term shoreline fluctuations caused by cross-shore transport are small compared to long-term changes the one-line model is believed to give a reasonable description of the shoreline evolution (Hanson and Larson 1987). Under these assumptions, the volume change per metre of shoreline can be calculated if the vertical elevation (i.e. closure depth to shoreface elevation above MSL) and the horizontal distance of the shoreline displacement are known. The volume (in m^3) can be calculated for a cell of specified beach length which is experiencing similar shoreline displacement (e.g. Drapeau and Mercier 1987).

(3) Beach Profile Surveys

Beach profiles respond to changes in incident waves and sediment input (Nordstrom and Jackson 1992). Repeated surveys of beach morphology allow assessment not only of complex process-response relationships (e.g. Wright and Short 1984; Nordstrom and Jackson 1992) but also of beach sediment budgets. Analysis of successive beach profiles surveys can be used to estimate beach volumetric change. This is often given as the volume change per metre of shoreline (i.e. m^3/m) (e.g. Brampton and Beven 1987; Savage and Birkemeier 1987; Lacey and Peck 1998).

Successive survey data can be used to quantify the change in a given beach cell by assuming that the change at one profile is representative over the distance between it and the half distance to each adjacent profile using:

$$\delta V = \frac{\delta A_j + \delta A_{(j+1)}}{2} L_{(j,j+1)} \quad (2.11)$$

where δA_j is the change in area at profile j (in m^3/m), δA_{j+1} is the change in area at the adjacent profile, and $L_{(j,j+1)}$ is the distance between the two profiles. Various studies have used this approach to quantify cell volume changes (e.g. Comber 1993; Drapeau and Mercier 1987; Deruig and Louise 1991; Foster et al. 1994; Hicks et al. 1999) using cell widths ($L_{(j,j+1)}$) ranging from ca. 100m (Foster et al. 1994) up to ca. 1km (Comber 1993;

Deruig and Louise 1991). The accuracy of the estimation of the volume change for a given cell is influenced by the cell width; smaller cell widths are likely to give more reliable estimates. To date, there has been no systematic analysis of successive surveys of closely spaced beach profiles in order to quantify the errors associated with volumetric gains and losses of sediment in a given beach cell depending on beach profile spacing. Indeed in the many studies that use successive beach profiles along a shoreline to estimate cell volume changes none quote the uncertainty of the estimates. Beach profile spacing must ensure that the profile is representative of the changes in the sediment cell it represents.

Definition of the closure depth for gravel sediment transfers

The depth of closure for a given or characteristic time interval can be defined as the shallowest depth seaward of which there is no significant change of bottom elevation and no significant net sediment exchange between the nearshore and the offshore (Wang and Davis 1999). Successive beach and bathymetric surveys can be used to define the depth of closure on sand beaches (e.g. Jimenez and Sanchez Arcilla 1993; Foster et al 1994; Wang and Davis 1999) or it can be calculated using empirical formulae based on extreme wave conditions (e.g. Hallermeier 1981; Birkemeier 1985). However, few studies have defined the depth of closure on gravel beaches and it remains an area of uncertainty. Diving experiments at Chesil beach identified a transition between mobile and immobile gravel occurring at a water depth of ca. 10m (Neate 1967). This is in agreement with results from wave refraction experiments at Mann Hill beach, Massachusetts which indicate that during normal wave conditions larger gravel cannot be moved at depths greater than 9m, although during severe storms (and thus high wave energy conditions) gravel can be moved in water depths of 19m (Brenninkmeyer and Nwankwo 1987). Kidson et al. (1958) and Neate (1967) demonstrate that gravel is normally mobile to depths of 6m under moderate wave heights and Comber (1993) suggested 6m is the maximum operational depth for gravel movement at Culbin, north-east Scotland.

Quantification of longshore transport rates using the concept of sediment continuity

Drapeau and Mercier (1987) and Jimenez and Sanchez Arcilla (1993) estimate longshore transport rates using actual beach cell volume changes and the concept of sediment continuity (i.e. equation 2.10) although neither study provides any estimate of the uncertainty in the transport rates.

Net longshore transport rates for the Ebro delta coast, Spain (Figure 2.7) were estimated from beach cell volume changes assuming a boundary condition of zero longshore

transport beyond the apex of the spit (i.e. $Q_o = 0$) (Jimenez and Sanchez Arcilla 1993). Transport rates were computed backwards using equation 2.10 and varied from 50 000 to 230 000 $\text{m}^3 \text{a}^{-1}$ along the coast. These agreed well with rates obtained using the CERC transport formulae and the average wave climate (Jimenez and Sanchez Arcilla 1993).

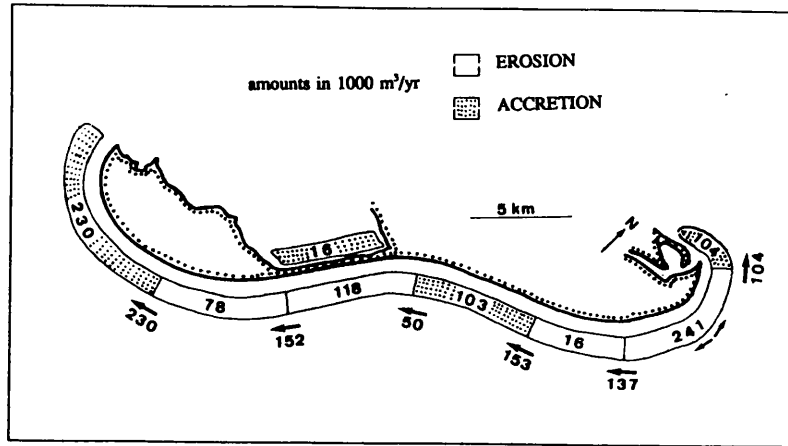


Figure 2.7: Sediment budget and net longshore transport rates in the southern hemidelta of the Ebro, Spain (source: Jimenez and Sanchez Arcilla 1993).

2.1.3 Fluvial-coastal sediment budgets

The sediment budget approach has been used extensively in fluvial environments (Section 2.1.1) and to a lesser extent in coastal environments (Section 2.1.2). To date there has been limited application of this approach to quantify sediment transfers from the fluvial to the coastal environment. Some examples are discussed below.

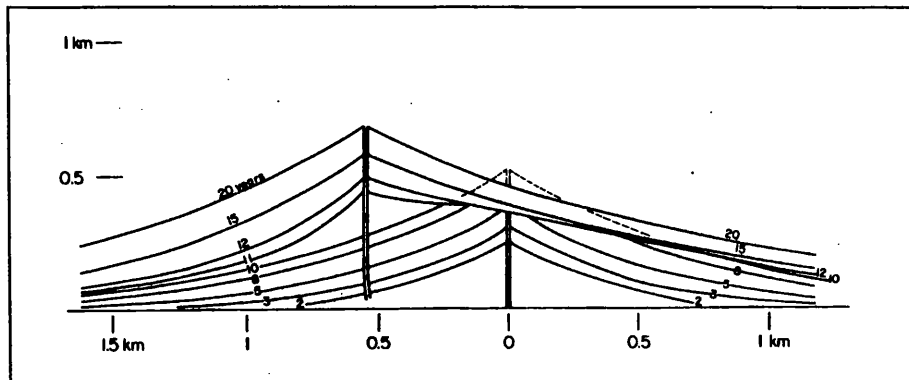


Figure 2.8: Numerical line model of a delta that consists of a complex pattern of shore-growth lines that could be represented in nature as beach ridges, resulting from a shift in the position of the river mouth after 10 years of delta growth. The shift results in the truncation of part of the original delta, followed by the overlap of the shorelines from the newly growing delta (source: Komar 1977).

Komar (1973, 1977) used the concept of sediment continuity (equation 2.8) to predict the evolution of a delta shoreline under various conditions of river sediment supply and wave parameters (Figure 2.8). This relatively simple numerical model used the basic principles of sediment budgeting to simulate characteristic river mouth and deltaic morphology and

predict the shoreline response to a fluctuating river mouth position (Figure 2.8). The shoreline growth lines can be imagined as representing a series of beach ridges; the complex morphology of which result from this simple shift of the river mouth as the shoreline adjusts to the change in sediment supply. This model could run backwards to provide a method for unravelling the history of development of beach ridge patterns at river mouths (Komar 1977).

A sediment budget was quantified for a 10km stretch of coast at the Rakaia river mouth, New Zealand (Kirk 1991) to determine the relative roles of marine and fluvial processes in the mouth system. Each term in the equation given in Figure 2.9 was estimated to give budget losses of between $270\,000\text{ m}^3\text{ a}^{-1}$ and $330\,000\text{ m}^3\text{ a}^{-1}$ for the system (Kirk 1991).

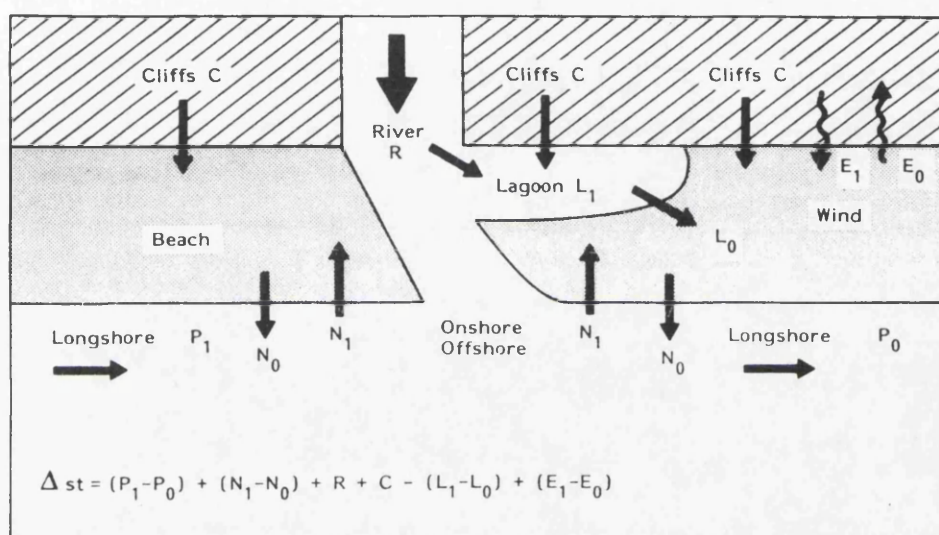


Figure 2.9: Schematic sediment budget and storage equation for a mixed sand-gravel river/beach/lagoon system. P_1 and P_2 represent the longshore drift into, and out of, the mouth region, respectively. N_1 and N_2 represent on and offshore transfers of sediment. R and C represent sediment inputs from the river and cliffs. Lagoon sedimentation is divided into storage (L_1) and losses to the coast (L_0). Onshore (E_0) and offshore (E_1) sediment transport by wind are not significant for coarse sediment transfers (source: Kirk 1991).

The negative sediment budget at the Rakaia suggests that marine processes dominate fluvial ones and the river output of coarse sediments is insufficient to maintain the coast and lagoon against sediment removal alongshore and thus long-term retreat (Kirk 1991). A similar study at the Waiau river mouth, New Zealand found that the supply of coarse sediment from the river was just enough to maintain the downdrift coast from erosion (Kirk and Shulmeister 1994). The coarse material load was derived primarily from erosion of previously stored sediment in the downstream reach of the river, due to recent flow regulation in the catchment causing lower mean flows and fewer floods (Kirk and

Shulmeister 1994). The river appeared to be cannibalising its stored sediment to maintain the coast and it was suggested that this supply will decline through time (Kirk and Shulmeister 1994).

A sediment budget for 140km of the Canterbury coast, New Zealand included the estimation of the gravel load of five rivers (Gibb and Adams 1982). Loads were estimated using an empirical equation derived from the Einstein-Brown formula (Adams 1980 cited in Gibb and Adams 1982) and combined with estimates of the gravel input from cliff erosion, abrasion rates and longshore transport rates to calculate the net north-easterly longshore transport rates (Figure 2.10). Estimated transport rates increase substantially at river mouths due to the input of sediment (Figure 2.10). Between the Rangitata and Rakaia Rivers the transport rate changes little as the gravel supplied by eroding cliffs roughly equals that lost by abrasion (Gibb and Adams 1982). Along the whole coast, abrasion accounts for 95% of the total gravel and sand input and less than 5% remains on the beaches in the north (Gibb and Adams 1982). No estimation of the uncertainties in the budget estimates were given.

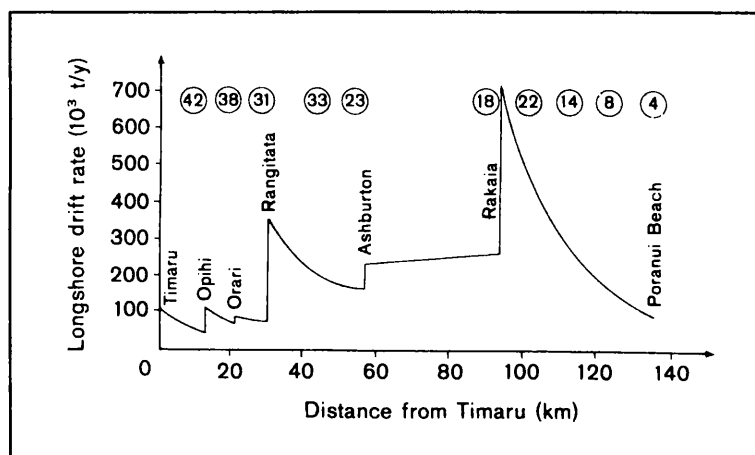


Figure 2.10: The net north-easterly longshore transport rate on the Canterbury coast calculated from longshore transport rates plus gravel input (cliffs are treated as line sources and rivers as point sources) less gravel lost by abrasion. The numbers above the graph are the percentage of gravel input from the south that remains on the beach (source: Gibb and Adams 1982).

The supply of sand from the San Lorenzo River and its dispersion along the Central Californian coast was estimated using a morphological approach with repeat river cross sections and beach profiles surveyed over a two year period (Hicks and Inman 1987). During winter floods, the sediment delivered to the coast was over ten times the mean annual supply and several times the mean annual longshore transport past the river mouth resulting in the formation an ephemeral delta (Hicks and Inman 1987). Paradoxically, the abundant sand supply from the river initially caused a temporary acceleration of erosion on

downdrift beaches, as the delta acted as a temporary groyne, interrupting the continuity of longshore transport (Hicks and Inman 1987). Again no estimation of the uncertainty in the fluvial transport rates and beach volume changes were given.

Using a sediment budget approach to estimate fluvial and coastal sediment transport rates requires the accurate estimation of reach storage changes (Section 2.1.1) and beach cell storage changes (Section 2.1.2). These are often obtained by repeat morphological surveys of the changing landform (i.e. cross-sections, beach profiles). However, as the temporal and spatial density of repeat surveys influences the accuracy of storage change estimates, budget studies must acknowledge all errors. This study proposes an adaptation of the methods used to quantify fluvial sediment transfers and coastal sediment transfers in order to quantify the transfers of gravel-sized sediment from rivers to the coast and its subsequent redistribution within the coastal zone.

2.2 Process environments

The sediment budget approach (Section 2.1) will be developed on the lower River Spey and the beaches of Spey Bay in north-east Scotland. The lower River Spey is a wandering gravel-bed river, which feeds sediment via a constantly changing spit/mouth complex to the gravel beaches of Spey Bay (see Chapter 3). The main characteristics of the three process environments which constitute the Spey system are outlined below.

2.2.1 Wandering gravel-bed rivers

The term 'wandering' is applied to gravel bed rivers that exhibit an irregular pattern of channel instability (Church 1983). Wandering rivers contain reaches that are essentially meandering and others that are braided, although the locations of these are not fixed through time, unless the river is constrained at particular locations. Meandering reaches are relatively stable and are areas of sediment transfer (Church and Jones 1982; Church 1983) whereas the braided reaches are laterally unstable, subject to avulsion and are locations of sediment storage. Sporadically mobile sediment is temporarily stored in these 'sedimentation zones' along the channel (Church 1983). These 'transfer' and 'sedimentation' reaches alternate and are of the order of 5 to 10 times the active channel width in length (Church and Jones 1982).

Wandering gravel-bed rivers migrate irregularly across their floodplains (e.g. Lewin and Weir 1977; Werritty and Ferguson 1980; Church 1983; Ferguson and Werritty 1983). They usually exhibit a zone of high activity (called the 'active channel') within the wider floodplain. Channel change and migration within the active channel is relatively frequent,

occurring during floods of 1 to 2 year recurrence interval. The boundaries of the active channel are also dynamic, and are continually modified (by erosion and deposition). Channel change which affects significant proportions of the remainder of the floodplain is less frequent, and occurs only during major flood events. The classification of sediment into discrete reservoirs (or stores) each with a *specific range of potential transfer conditions* is a useful approach given the above characteristics of wandering gravel-bed rivers (e.g. Kelsey et al. 1987, Figures 2.1 and 2.2).

The channel pattern found in wandering rivers varies through time as a result of the impacts of flood events of different size. Given the rather random magnitude and frequency of competent floods, it is suggested that channel changes in wandering rivers are stochastic in nature, perhaps following some type of Markovian process (Ferguson and Werritty 1983; Kelsey et al. 1987). The channel pattern of the River Feshie becomes more complex (braided) after high magnitude flood events, gradually simplifying during more moderate events (Werritty and Ferguson 1980; Ferguson and Werritty 1983). The complex morphology of wandering rivers, created by the passage of numerous floods of varying duration and magnitude is reflected in the bar sedimentary structures, which exhibit spatial and temporal variability from fining-up deposition (during falling stages) and coarsening upwards deposition (during the rising stage of floods) (Ferguson and Werritty 1983).

Spatial and temporal variability in bedload transport rates and storage volumes have long been recognised, both in the field (e.g. Gilbert 1917; Griffiths 1979; Church 1983; Ergenzinger 1988) and in laboratory flume studies (e.g. Ashmore 1987; Hoey and Sutherland 1991). The term 'bedload pulse' is used to describe the periodicity in bedload transport rates at a particular site (Hoey 1992). The spatial manifestation of the passage of a pulse is termed a 'bed wave' (Hoey 1992), which is an increase in sediment storage in a reach, relative either to that reach at preceding or succeeding times or to adjacent upstream and downstream reaches at the same time. It has been argued that macro-scale spatial and temporal fluctuations in sediment transport and storage volumes are inherent features of braided and wandering rivers (Davies 1987; Goff and Ashmore 1994; Nicholas et al. 1995) and are the response to complex interactions between upstream sediment supply and discharge, the relative timing of which is critical (Lane et al. 1996). Hoey (1992) and Goff and Ashmore (1994) describe phases of erosion and deposition which are independent of changes in discharge and are primarily associated with variations in upstream sediment supply. It is argued that sediment transport in the braided Ohau River, New Zealand is supply limited and not flow limited (Davoren and Mosley 1986) as measurements indicate

that there is no unique relationship between bedload transport rates and hydraulic conditions. However, cycles of disturbance and recovery in response to flood events (e.g. Ferguson and Werritty 1983) highlight the important control of discharge. The impact of sediment supply and discharge events on a particular river is dependant on their relative timing, order, inter-arrival time and absolute magnitude.

Gravel-bed rivers typically contain a wide range of grain sizes in their bed, bank and bar material (Ashworth et al. 1992b). Sediment is selectively sorted (by size, shape and weight) to produce the depositional structures common in gravel-bed rivers: downstream fining of sediment is nearly ubiquitous (e.g. Church and Kellerhals 1978; Morris and Williams 1999); barhead-to-tail fining is commonly noted (e.g. Bluck 1982); and vertical armouring is common (Andrews and Parker 1987). The initial size mix, sediment supply, flow hydraulics and channel pattern all interact to influence the type and degree of bed-material sorting that occurs in gravel bed rivers (Ashworth et al. 1992a). The formation of a coarse surface layer can significantly influence bedload transport rates (e.g. Davoren and Mosley 1986; Andrews and Parker 1987). Lower flows are unable to break the armour layer, transporting limited quantities of sediment. When a flood flow capable of breaking up the armour layer occurs, a significant rise in the quantity of bed material transferred downstream is observed. The size distribution of the subsurface material (i.e. beneath the coarse armour) is similar to that of the long-term averaged bedload and is characteristic of the bulk of the sediment stored in a river reach (Andrews and Parker 1987).

2.2.2 River mouths

Rivers provide an important supply of sediment to the coastal zone (Table 2.1). Deltas are subaerial and submarine protuberances extending out from shorelines in situations where sediment is supplied to the coastal zone more rapidly than it can be redistributed by coastal processes (Zenkovitch 1967). Deposition of the fluvial bedload occurs at the delta because of the radial outflow of decelerating river water and generally occurs over a short distance.

Delta morphology depends on the river discharge, sediment load and the wave regime of the coast (Wright and Coleman 1973). The relative dominance, magnitude and frequency of the fluvial and marine 'signals' determines the geomorphic expression at the mouth. If fluvial processes dominate and sediment load is high the delta will build out as a protuberance in the shoreline. If wave processes dominate the sediment input is likely to be rapidly redistributed in the coastal zone and there is likely to be only a small protuberance in the shoreline marking the location of the river mouth. The relationship between fluvial

and marine processes thus influences the amount and permanence of sediment storage at the delta.

Ternary diagrams, such as those proposed by Galloway (1975), Wright (1985) and Boyd et al. (1992) utilise the relative importance of river, wave and tide power to classify river mouths; the extremes being river, wave and tide-dominated with a full range of intermediate types (Figure 2.11). Delta morphology is also influenced by the stability and channel pattern of the fluvial supply, sediment load and grain size, channel gradient, flow velocity and the distance of the shoreline from the source (Figure 2.12). Fan and braid deltas are coarse-grained and contrast in shape, size and composition with fine-grained deltas (McPherson et al. 1987).

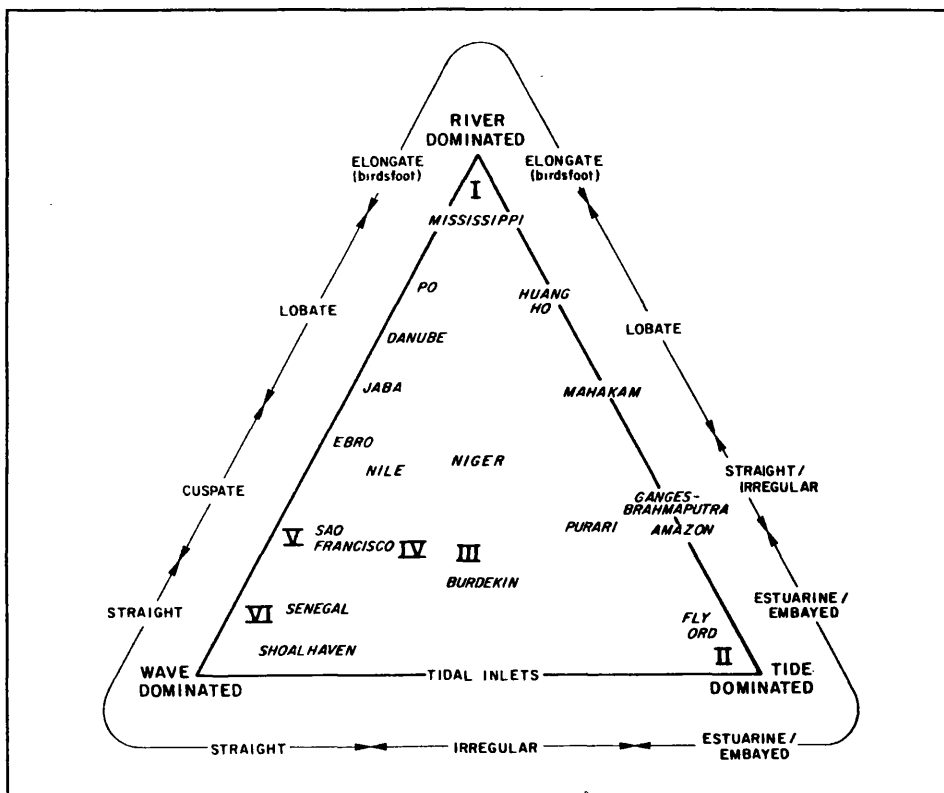


Figure 2.11: Ternary classification of deltas depending on the relative importance of river, wave and tidal processes showing typical delta shapes (source: Wright 1985).

A classification of river mouths, based on the relative influence of river-derived sediment load on the stability trend (erosion or accretion) of the adjacent coastline was advanced by Zenkovitch (1967) who made a fundamental distinction between 'large' and 'small' rivers. 'Large' rivers contribute abundant sediment load to the coast so that it either maintains a stable position against losses due to abrasion and longshore transport, or it actively accretes. In contrast, 'small' rivers produce insufficient sediment load to protect the coast from direct marine erosion and storm attack. The terms 'large' and 'small' are thus relative with respect to the receiving coast (Zenkovich 1967) and as Kirk (1991) notes, the key

factor is not the total river sediment load, but the proportion of the (bed) load coarse enough to nourish the receiving coast.

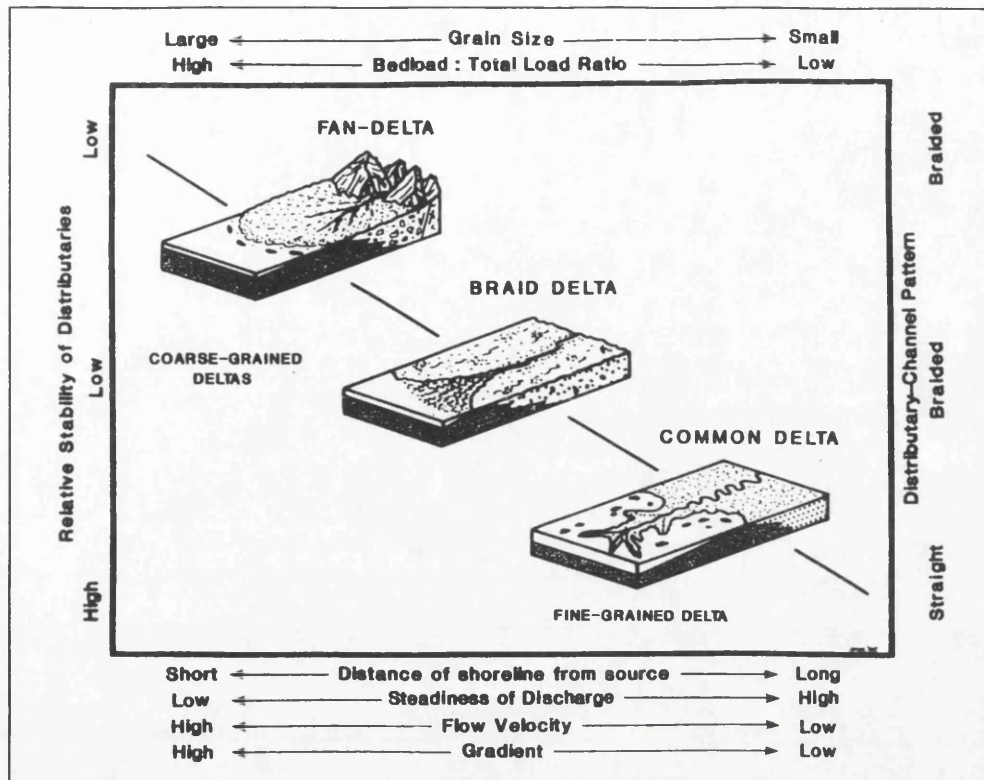


Figure 2.12: A comparison of fan deltas, braid deltas and fine-grained deltas based on distributary channel patterns and stability, sediment load and size, stream gradient and velocity and distance of shoreline from source (source: McPherson et al. 1987).

Zenkovitch (1967) noted that research on small rivers or the marine forces at their mouths was limited, with most literature concerned with large rivers (particularly deltas) and fluvial processes. Fluvial-coastal interactions remain extensively studied on deltaic coasts (e.g. papers in Colella and Prior 1990; Suter 1994; Allison 1998; Eisma 1998), however despite recent research in New Zealand (Kirk 1991; Shulmeister and Kirk 1993; Kirk and Shulmeister 1994) and elsewhere (Hicks and Inman 1987; Jimenez et al. 1997) fluvial-coastal interactions on small rivers remain relatively poorly understood (Shulmeister and Kirk 1997). Further research is required to fully understand the storage and transfers of sediment, marine forces and the sediment budgets at small river mouths.

The morphology at the mouths of 'small' rivers on wave dominated coasts often display features such as spits and bars (e.g. Kidson 1963; Kirk 1991) since wave processes dominate and redistribute the fluvial input of sediment. Longshore currents extend the spit in the downdrift direction, leading to the deflection of the river mouth (e.g. Kirk 1991; Shulmeister and Kirk 1993, 1997). The downdrift extension of spits often requires a supply of sediment from the updrift section of the coast, which may lead to erosion of the updrift part of the spit or the updrift beach. Spits may be breached during high river flows (e.g. Grove

1955; Kirk 1991) and may in some cases become detached from their source river to migrate downcoast (e.g. the Findhorn, north-east Scotland (Hansom 1999), Figure 2.13)

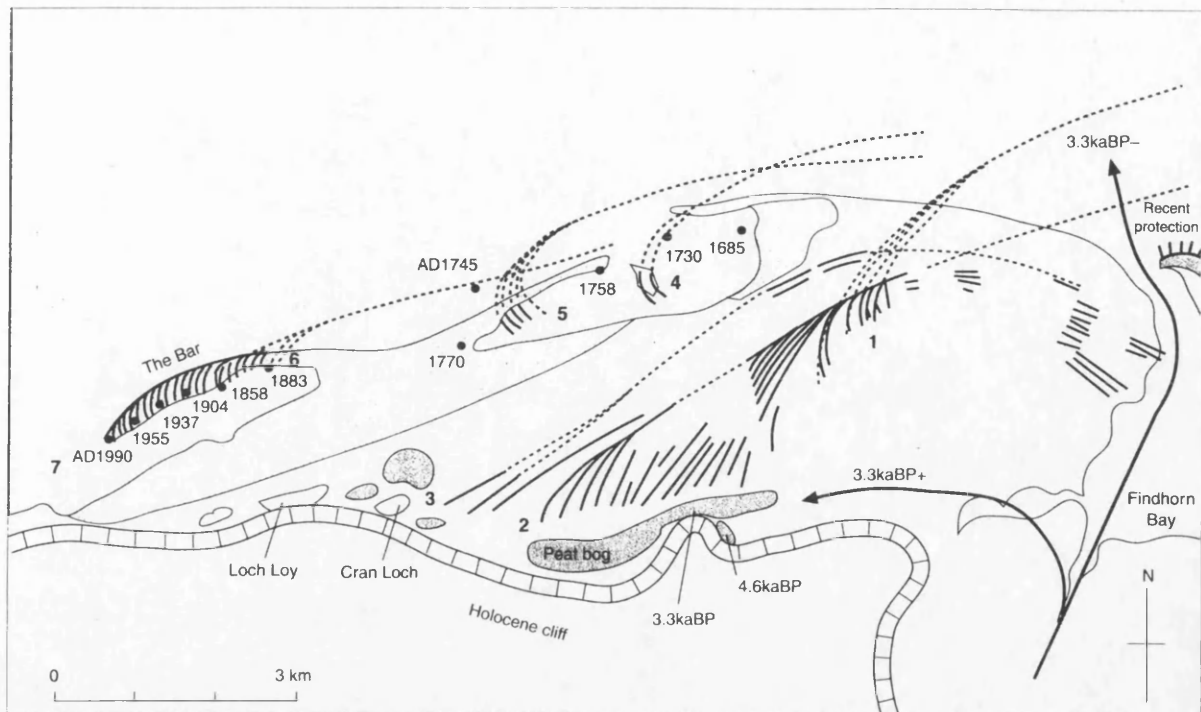


Figure 2.13: The downdrift extension of spits at the mouth of the River Findhorn, North-east Scotland. The coastal system has developed by updrift erosion fuelling downdrift accretion. The distal end of the present spit (The Bar) is now detached from its source and migrates downcoast (source: Hansom 1999).

Fluvial-coastal interaction can cause large-scale irregular variations in the regional longshore transport regime, leading to temporal and spatial variations in the rate of coastal erosion along the downdrift coast (Hicks and Inman 1987; Kirk 1991; Hicks et al. 1999). For example, the Rakaia river promotes spit development and extension during moderate to low flows, while spit breaching occurs during high flows. Thus, during spit elongation, the lagoon enclosed to the landward acts as a *sink* of sediment, whereas during and immediately after high river flows it acts as a significant source of coastal sediment (Kirk 1991). When breaching occurs a 'slug' of sediments, temporarily stored in the spit and lagoon, is injected into the downdrift coast. During spit elongation, the temporary storage of sediment in the spit and lagoon can induce starvation of downdrift shores and so temporarily accelerate erosion there (Kirk 1991). The processes of spit elongation and natural breaching occur at the mouth of the River Findhorn (Figure 2.13) and the Spey (Grove 1955; Omand 1976).

Sediment bypass (or storage) processes at river mouths are complex and vary depending on both the geomorphic setting and the sediment characteristics of the drift material. There is

ample evidence that sand bypasses river mouths (e.g. Bruun and Gerritson 1959); this occurs either by bar bypassing in the surf and nearshore zone, or by tidal bypassing involving tidal entrainment first into and then out of the inlet, or by a combination of both (Kirk 1991). Kidson et al. (1958), Kidson (1963) and Carr (1965) provide examples of the passage of gravels across river mouths, although in many cases only part of the drift will bypass, the remainder being temporarily deposited in large spits or bars. At the Rakaia river mouth, most of the sediment travelling as beach drift bypasses in the process of the spit growth-breach sequence (see above) and is given the term ‘spit bypassing’ (Kirk 1991). It is argued that this process creates pulses in the longshore transport rate which have been suggested to occur on the South Canterbury coast, New Zealand (e.g. Neale 1987 cited in Kirk 1991; Todd 1989; Single and Hemmingsen 2000).

The supply of gravel sediment from rivers to the coast is a notoriously difficult variable to quantify (Crofts 1974; Kirk 1991) and is often episodic in response to large floods (Gibb and Adams 1982). Sediment yields were calculated for all New Zealand rivers by Griffiths and Glasby (1985) assuming that bedload (i.e. coarse sediment load) is ca. 3% of the suspended load (Table 2.2). The annual sediment yield for the River Spey was estimated as $3.3 \times 10^4 \text{ t yr}^{-1}$ by Reid and McManus (1987) assuming that the bedload comprised 10% of the suspended load. New Zealand sediment yields are considerably higher than those of the Spey (Table 2.2).

Catchment	Area (km ²)	Channel slope (m m ⁻¹)	Mean rainfall (m yr ⁻¹)	Mean discharge (m ³ s ⁻¹)	10-year flood discharge (m ³ s ⁻¹)	Sediment yield (t yr ⁻¹)
Waimakariri	3210	0.006	1.90	120	2 708	5.3×10^6
Rakaia	2640	0.01	3.00	200	3 764	4.3×10^6
Ashburton	540	0.01	1.40	8	170	3.1×10^5
Spey	3011	0.004	0.93	64	1 100	3.3×10^4

Table 2.2: Hydrological characteristics and sediment yields from a selection of New Zealand rivers and the River Spey, north-east Scotland (source: compiled from Griffiths and Glasby 1985, Kirk 1991 and Reid and McManus 1987).

While the above research has advanced current understandings of the morpho-dynamics of fluvial-coastal interactions at ‘small’ river mouths, there remain several areas of uncertainty. Sediment exchange processes at river mouths remain poorly understood, in particular the mechanisms of spit growth and river mouth deflection which may cause spatial and temporal variations in the longshore transport rate require further investigation. Further understanding of the interaction of wave and fluvial processes and their

relationships with sediment exchanges at river mouths is required. In addition, there is little quantitative information about the supply of gravel-size sediment from rivers to the coast. To date, all estimates of gravel transport to coasts have been based either on formulae (e.g. Gibb and Adams 1982) or from estimates of the total sediment yield (e.g. Kirk and Hewson 1979; Kirk 1991). There has been no field estimation of the transfer of gravel-sized sediment from rivers to the coast, although the morphological approach has been used to quantify sand transfers to the coast (Hicks and Inman 1987). This study aims to advance current knowledge of these uncertainties.

2.2.3 Coarse-clastic beaches

Following Carter and Orford's (1993) review of the somewhat perplexing terminology commonly used in the coarse-clastic beach literature (e.g. pebble, cobble, boulder, shingle, gravel, blocks etc.) the term coarse-clastic is adopted here. This term covers the full spectrum of textural properties of beach sediment, although its use in the context of beaches and barriers has been restricted mainly to the larger end of the grade size scale (Carter and Orford 1993). The term gravel is used to cover the range of sediment sizes from pebble to boulder (2 - 1024mm diameter) of which coarse-clastic beaches and barriers are commonly formed (Carter and Orford 1993).

Coarse-clastic shorelines are found throughout the world, but particularly on mid- and high-latitude coasts situated in formerly glaciated regions where the supply of coarse sediment is, or has been, plentiful. The terms 'paraglacial' and 'quasi-paraglacial' have recently been introduced to describe such coasts (Forbes et al. 1995). Paraglacial coasts are those which have developed on or adjacent to formerly ice-covered terrain, where glacial landforms or glacial sediments have a strong influence on the nature and evolution of the coast and quasi-paraglacial coasts are those dominated by fluvial deposits incorporating glacial outwash (Forbes et al. 1995).

Coarse clastic beaches have been studied extensively world-wide (e.g. Bluck 1967; Kirk 1980; Carter and Orford 1984; Forbes et al. 1995; Bird 1996; Orford et al. 1996; Bartholoma et al. 1998). From this work, the distinctive morphosedimentary and morphodynamic characteristics of coarse clastic beach and barrier environments can be summarised and typically include (adapted from Forbes et al. 1995):

- steep and reflective beach-face slopes (often with low-angle platforms or aprons at the base, which may be formed of soft (muds, sands) or hard (gravel, cohesive clays, rock) substrates). The presence of these low-angle platforms or aprons adds a dissipative

element into a predominantly reflective environment (Bluck 1967; Carter and Orford 1993);

- high permeability, swash infiltration and seepage potential (Kirk 1991; Forbes et al. 1995);
- high entrainment thresholds and hydrodynamic roughness (except where large clasts move across a finer substrate) (Isla and Bujalesky 1993; Forbes et al. 1995);
- particle shape and size interaction in sediment transport and sorting (Bluck 1967; Bird 1996); and
- restricted influence of wind and vegetation effects (in the absence of significant sand volumes for dune construction) (Kirk 1991; Forbes et al. 1995).

Steep beach-face slopes and the generally reflective nature (as defined by Short 1979) of coarse clastic beaches (Carter and Orford 1993; Forbes et al. 1995) leads to the development of low mode harmonic and sub-harmonic edge waves in the nearshore. As a result suites of cusps of different wavelengths on the beach-face are common, but ephemeral, features on coarse clastic beaches (Sherman et al. 1993). Cusp morphology and the associated shape and size sorting may exert a strong control over subsequent sediment movement and can act as a template for other processes, most notably the occurrence of overwash during storms (Orford and Carter 1984).

Most coarse clastic beaches comprise varying mixtures of gravel and sand (Carter and Orford 1984; Carter and Orford 1993). The relative proportion of gravel to sand has a controlling influence on beach morphology and resulting sediment transport processes (McLean and Kirk 1969; Kirk 1980). Cross-shore separation of sands onto the lower foreshore and nearshore aprons is extremely common in mixed sand and gravel beaches (Kirk 1980; Carter and Orford 1993; Forbes et al. 1995).

Coarse-clastic beaches and barriers display considerable variety in form (Figure 2.14). The coarse-clastic beaches of north-east Scotland (Ritchie et al. 1978) display similar forms to the Irish and Scottish examples in Figure 2.14, but also display some characteristics of the New Zealand mixed sand and gravel beaches described by Kirk 1980 (Figure 2.15).

A fundamental textural distinction is noted on the mixed sand and gravel (MSG) beaches of New Zealand (Figure 2.15) between the beach deposits, which contain a wide size range of predominantly coarse material, and those in the nearshore, which comprises a much narrower range of predominantly finer material (Kirk 1980). From this, it is implied that *transfers* of sediment on the beaches and in the nearshore involve different size ranges of

sediment moving in distinctly different and quite separate transport systems. It follows that, on MSG beaches which display a distinct cross-shore separation, there is no periodic onshore-offshore re-circulation of sediment between the subaerial beach face and the nearshore sea-bed as is characteristic of most sand beaches and some pure gravel beaches (Kirk 1980).

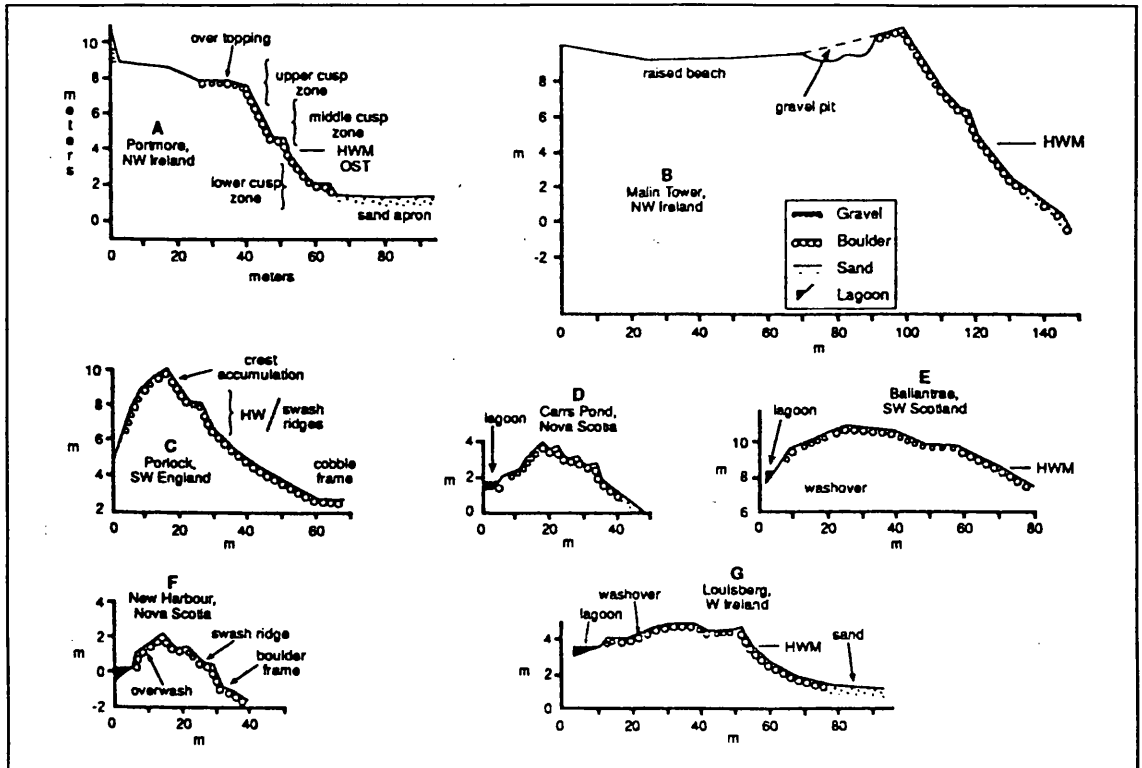


Figure 2.14: Coarse-clastic barrier and beach cross-profiles drawn to a common scale to illustrate the variability in form (source: Carter and Orford 1993)

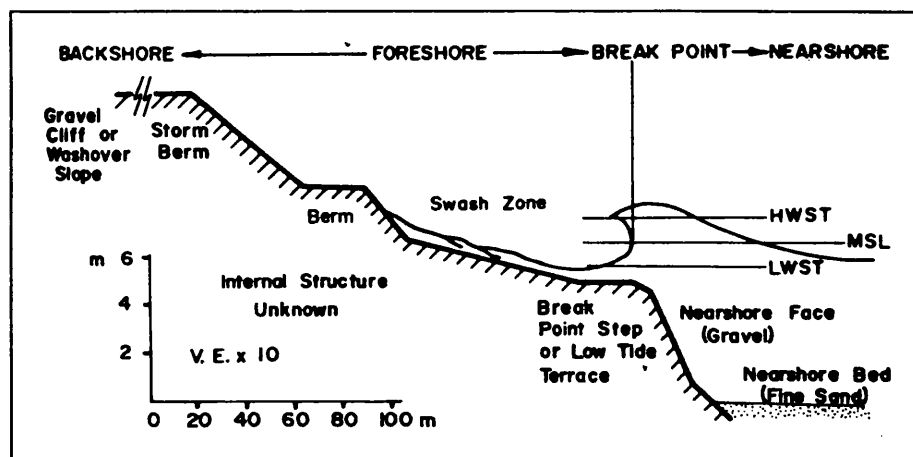


Figure 2.15: Typical morphology and zonation of mixed sand and gravel beach profiles (source: Kirk 1980).

Sediment transport studies on coarse-clastic beaches are sparse and inconsistent compared with their sand counterparts (Bray 1997); this is a result of their relative global scarcity

(Carter and Orford 1984) and the technical difficulties of undertaking the necessary field measurements, especially during storms (Bray et al. 1996).

On coarse-clastic shorelines, sediment transport is generally concentrated within a narrow zone between the breakers and the beach face and is dominated by bedload transport (Carter and Orford 1993) although results from tracer experiments indicate that gravel transport is most rapid on the upper beach near the high water mark (Bray 1997). Isla and Bujalesky (1993) note that there is an important population of the coarser and more spherical clasts saltating. The general nature of the bed must be taken into account when studying coarse clastic sediment transport (Carter and Orford 1993). Long-shore and cross-shore sorting on gravel beaches (e.g. Bluck 1967; Carr 1971; Williams and Caldwell 1988; Bird 1996) can lead to beaches developing an organisational framework, with diagnostic cross-shore and along-shore facies assemblages. Widespread armouring (Isla 1993), imbrication (Carter and Orford 1993) and sorting of coarse-clastic beaches can result in the stabilisation of the bed (in a similar way to river beds, cf. Andrews and Parker 1987).

The standard CERC (1984) model which relates longshore sediment transport (Q_s) to longshore energy flux (P_L) is well developed for sand beaches, with a value of 0.77 derived for k (CERC 1984; Komar 1990):

$$Q_s = k \cdot P_L \quad (2.12)$$

where:

$$P_L = (EC_n)_b \sin \alpha_b \cos \alpha_b \quad (2.13)$$

and $(EC_n)_b$ is the wave-energy flux or power evaluated at the breaker zone and α_b is the wave-breaker angle.

The model was originally derived for sand beaches and the problems of applying such formulae to coarse clastic beaches are recognised (e.g. Carter and Orford 1993). There have been several studies of gravel transport on beaches with the principal aim to estimate k for coarse clastic beaches (e.g. Wright et al. 1978; Brampton and Motyka 1987; Hattori and Suzuki 1987; Nicholls and Wright 1991; Bray et al. 1996). All studies report much lower values of k for gravel transport compared to sand transport (e.g. 0.0025 (Hattori and Suzuki 1987), 0.002 (Brampton and Motyka 1987)). Results from numerous gravel tracing experiments in southern England report values of k between 7 and 100 times lower than those for sand (Nicholls and Wright 1991), implying that gravel transport is much less efficient than sand. This is predominately a function of grain size and sorting (Nicholls and

Wright 1991), although armouring and imbrication of clasts may also be important (Carter and Orford 1993; Isla 1993).

However, recent work on gravel transport *during high energy conditions* suggest that the transport efficiency increases by an order of magnitude during storm events to approach that of sand (Bray et al. 1996). Gravel drift volumes vary significantly according to wave energy, with rapid bursts of drift (averaging $3\ 300\ \text{m}^3\text{tide}^{-1}$) generated by storm events at Shoreham beach, southern England (Bray et al. 1996). Results from tracer experiments in New Zealand led Matthews (1980) to suggest that beach gravels move alongshore as small slugs near low water level during periods of high wave energy. During normal low wave energy conditions the slugs of gravel are welded smoothly onto the berm, retaining no morphological expression (Matthews 1980). Brunsden (1999) argued that slugs of pebbles move along Chesil Beach, Dorset in response to storm events, although the model he proposes is difficult to demonstrate. Temporal and spatial variations in longshore gravel transport can occur in response to high wave energy conditions (Bray et al. 1996; Brunsden 1999) and river mouth processes which cause variations in local sediment supply (Hicks and Inman 1987; Kirk 1991).

Current understanding of gravel transport on beaches is relatively limited compared to sand. In particular the nature and morphological expression of gravel transport requires further investigation.

2.3 Summary and research approach

This chapter has critically reviewed the sediment budget literature and assessed its applicability to estimate transport rates in fluvial, river mouth and coastal systems. The sediment budget approach is useful technique for estimating sediment transport rates in wandering gravel-bed rivers (Ashmore and Church 1999) although an estimation of the uncertainty in the transport rate estimate should always be given (cf. Martin and Church 1995). This has often been omitted in previous studies (e.g. Griffiths 1979; Ferguson and Ashworth 1992). Estimation of transport rates in the coastal system using the sediment budget approach is less well developed, although several studies (e.g. Gibb and Adams 1982; Drapeau and Mercier 1987; Jimenez and Sanchez Arcilla 1993) have advanced the technique. Again there has been a tendency to ignore any uncertainties in the transport rate estimates.

This study will develop the sediment budget approach to quantify sediment transport rates from fluvial to coastal systems. The research is largely field-based and will utilise repeat

morphological surveys of cross-sections, beach profiles and repeat planimetric maps to estimate transport rates on the lower River Spey and the beaches of Spey Bay. Uncertainties will be estimated and the applicability of each approach to estimate transport rates (i.e. cross-sections, reach scale surveys, beach profiles and planimetric mapping) will be evaluated. It is hoped that some of the uncertainties concerning river mouth dynamics (Section 2.2.2) and longshore gravel transport dynamics (Section 2.2.3) will be advanced by this study.

3. THE FIELD SITE

The research aims set out in Chapter 1 require the identification of a dynamic site which permits the natural transfer of gravel-sized sediment from the fluvial system to the coast. The lower River Spey as it enters the Moray Firth at Spey Bay, north-east Scotland (Figure 3.1) fits this requirement. The site incorporates the active, wandering, gravel-bed reach of the River Spey ca. 3km upstream of the mouth, the active and constantly changing delta complex at the river mouth and the mainly coarse-clastic shoreline of Spey Bay. The harbour/sea-wall of Porttannachy and the harbour/cliffs of Lossiemouth mark the eastern and western boundaries of both the site and the littoral sediment cell, the definition of which is a necessary prerequisite for sediment budget calculations (Komar 1996) (Section 2.1.2). The Spey Bay sediment cell is bounded by the rock coastline at Portknockie in the east and the headland at Branderburgh in the west, both of which act as drift divides (H.R. Wallingford 1995). The geology, geomorphology and process environments of the field site are introduced in this chapter.

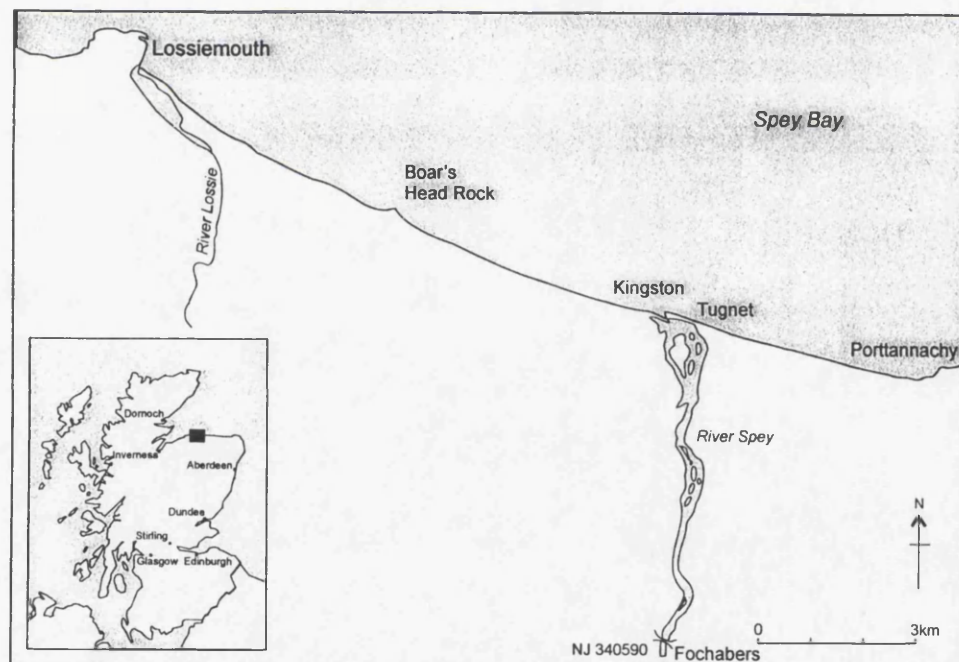


Figure 3.1: Location map of the lower River Spey and Spey Bay, north-east Scotland

3.1 Geology of the Moray Firth

The Moray Firth is a large Mesozoic basin (Figure 3.2) with a broadly conformable succession from Devonian to Cretaceous dipping uniformly towards the centre where the thickness of the succession increases substantially (Chesher and Lawson 1983). Onshore,

the Dalradian series in the east of the Moray Firth gives way towards the west to the Moinian rocks of the Inverness area, with granitic intrusives appearing at a number of localities (Robertson 1990). Along the southern coast of the Moray Firth the basement is unconformably overlain by the sandstones and shales of the upper Devonian Old Red Sandstone (Figure 3.2). Further details of the solid geology can be found in Gemmell et al. (2000).

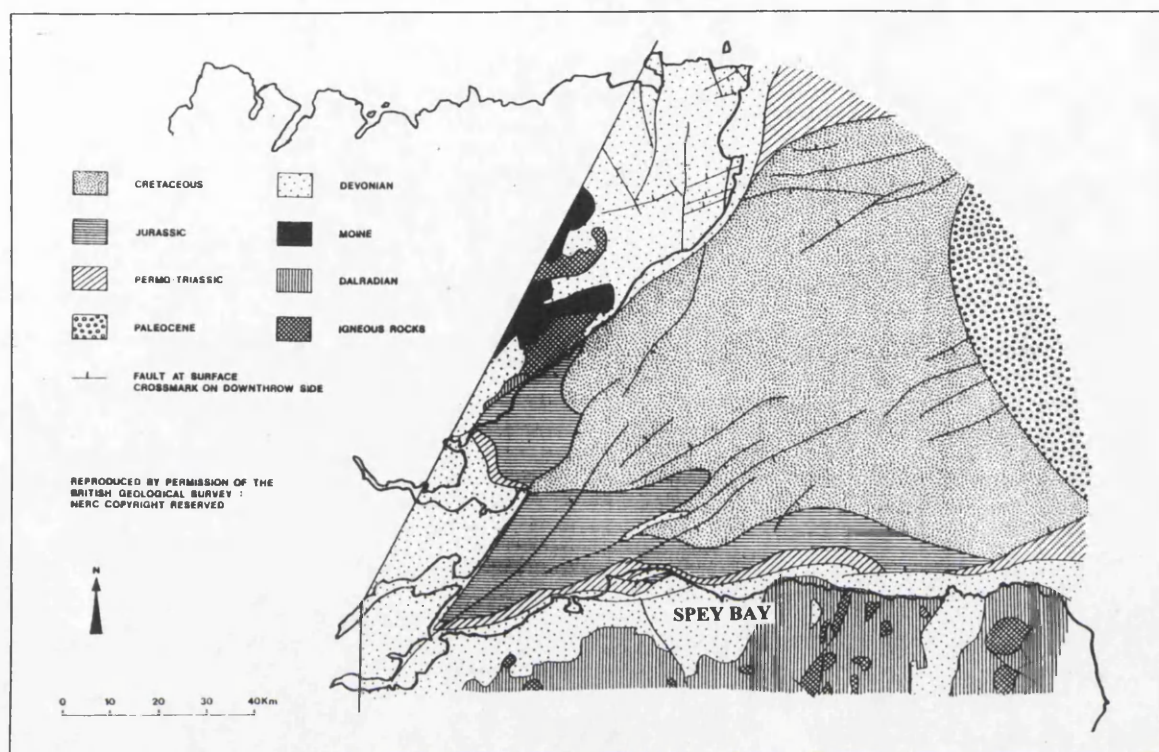


Figure 3.2: Solid geology of the Moray Firth (source: adapted from Andrews et al. 1990)

Quaternary deposits cover almost the entire Moray Firth area, reaching depths of up to ca. 70m (Figure 3.3). Chesher and Lawson (1983) subdivided the Quaternary deposits into northern and southern units based upon thickness. The northern units are poorly defined, generally thinner deposits with varied accumulation sequences. The southern units which most affect the Spey Bay area were found to be much thicker, and have been further subdivided into a series of five elongate sediment-filled basins aligned approximately E-W (Figure 3.3). For example, within the south Lossiemouth basin, 5km off the Spey Bay coast, a borehole penetrated a 27m thick pocket of sand and gravelly sediment overlying Permo-Triassic sandstone (for detailed stratigraphic details see Chesher and Lawson 1983).

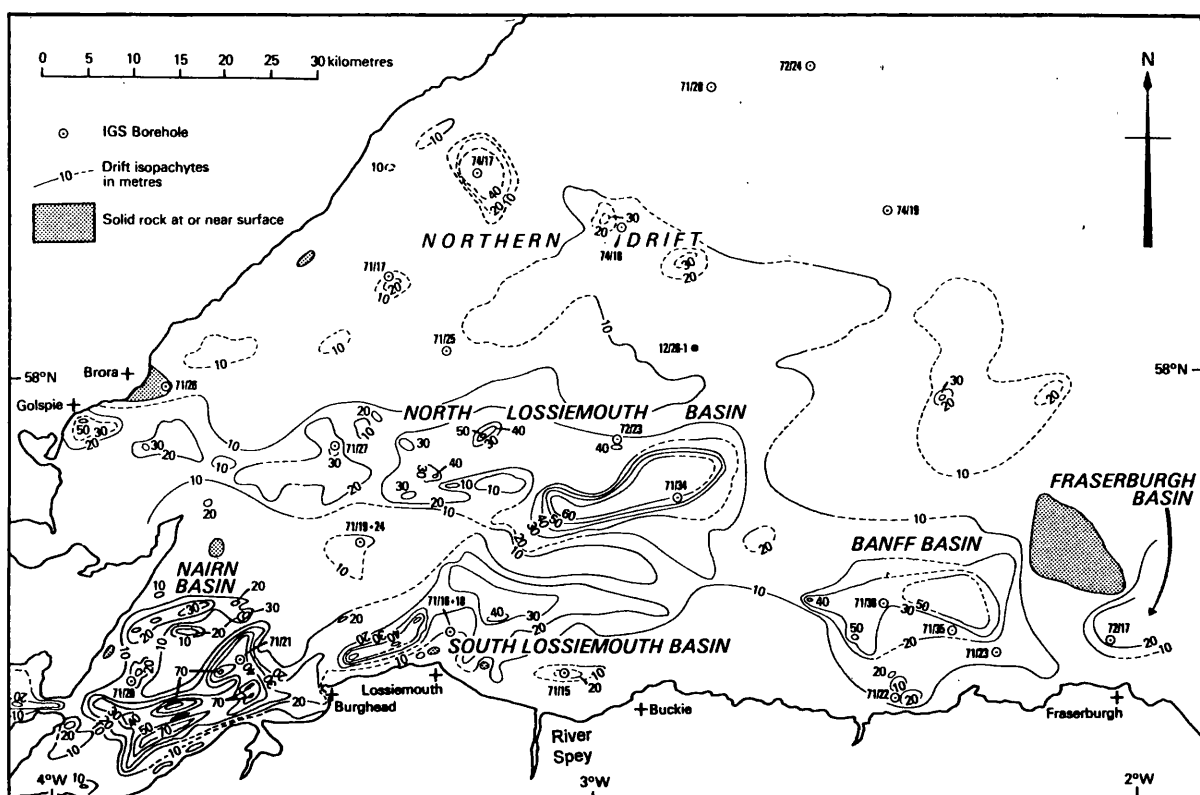


Figure 3.3: Quaternary basins of the Moray Firth (source: Chesher and Lawson 1983)

3.2 Late Quaternary sea level history of the Moray Firth

Final downwasting of Moray Firth ice at ca. 13000 BP allowed flooding of newly deglaciated areas by the sea (Firth 1989). However, rapid isostatic recovery of the land surface during this period outstripped the rate of eustatic sea level rise, producing a fall in relative sea level (RSL) which was thought to be already low by the onset of the Loch Lomond Stadial (ca. 11-10000 BP) (Synge 1977; Haggart 1986, 1987).

A minimum age for the fall in RSL prior to the onset of the Holocene transgression in the inner Moray Firth is 9610 ± 130 BP (Haggart 1986, 1987). Further evidence for a low sea level at this time is provided by a series of extensive intertidal peat deposits found below HWST around the Moray Firth. The end of this period has not been identified, but Firth and Haggart (1989) record a falling RSL ca. 9200 BP, and Peacock et al. (1980) recorded a possible low stand at -6m OD dated at 8748 ± 100 BP in the Cromarty Firth.

This period of falling RSL was reversed by a major rise in eustatic sea level (the Holocene Transgression) (Fairbanks 1989). Haggart (1986) suggested that RSL was rising by ca. 8800 BP in the Beaully Firth and Firth and Haggart (1989) dated the culmination of this rise to ca. 6400 BP. It was marked in the Beaully and inner Moray Firths by the formation of the

Main Postglacial Shoreline (MPS) and a series of raised gravel ridges up to ca. 9m OD at Spey Bay.

Since the peak of the Holocene Transgression, RSL has displayed a falling trend to the present (Firth and Haggart 1989), as a result of continued isostatic recovery, coupled with a reduction in the rate of eustatic sea level rise. A series of raised shoreline features has been identified around the inner Moray Firth at successively lower altitudes below the MPS, suggesting minor stillstand events within the overall scenario of falling RSL to the present (Firth and Haggart 1989). The Late Quaternary sea level history for Spey Bay is summarised in Figure 3.4.

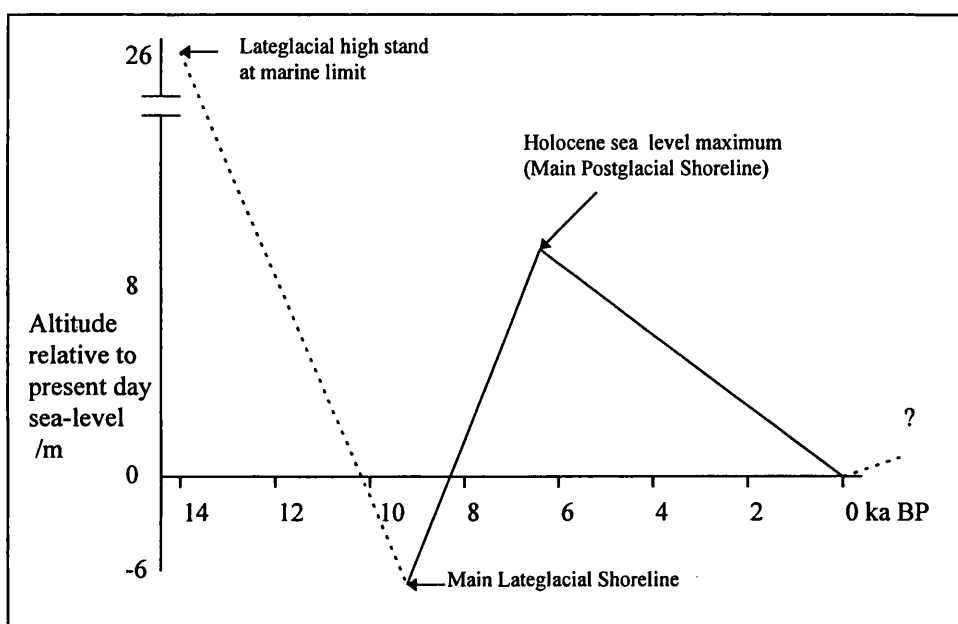


Figure 3.4: Diagrammatic relative sea-level curve for Spey Bay (source: Gemmell et al. 2000)

Holocene Sediment Supply

Decay of late Devensian ice occurred as climate ameliorated and by ca. 13000 BP Scotland was largely ice free (Sutherland 1984). The local Moray Firth glacier retreated towards the west, releasing vast amounts of clastic sediment from its snout. Similar decay of ice in the Cairngorms to the south released large volumes of sediment-laden meltwater to the Findhorn and Spey, at discharges considerably higher than experienced currently (Young 1977; Maizels 1988). The low RSL at this time allowed subaerial sedimentation across the inner continental shelf. RSL continued to fall (Firth and Haggart 1989) to a low stand at ca. -6m OD (Peacock et al. 1980) before rising again to a high stand ca. 6500 BP at the peak of the Holocene Transgression (Figure 3.4). The effect of this rise in RSL was to carry sediment onshore from the inner shelf, creating a sediment-rich coastal environment. Additionally with RSL at a higher level than at present (ca. +8m OD), the flooded basin,

known today as the Laigh of Moray south of the high ground of Burghead/Lossiemouth, created a marine corridor south of an offshore island (Comber 1993).

Under conditions of net westerly drift, it is likely that gravels from the Spey moved freely alongshore through this corridor, into a proto-Burghead Bay and west to Findhorn Bay (Comber 1993). Combined with the net onshore movement of sediment under a rising RSL, a strongly positive sediment budget was created within Spey and Burghead Bays (Comber 1993). In situations where sediment supply is plentiful, storm ridge deposition tends to result in sequential additions of further ridges on the seaward edge and to develop a pattern of multiple sub-parallel ridge deposition (Carter et al. 1987). This pattern can be seen today at Spey Bay (with gravel ridges 800m wide over ca. 15km of coast) and at Culbin (gravel ridges 4km wide over ca. 7km of coast) (Comber 1993, Hansom 1999, 2000).

Under the falling RSL that occurred post 6500 BP, water depths in the Laigh of Moray/Loch Spynie channel were reduced and, in association with substantial sediment deposition, the corridor gradually became choked with westwards drifting sediment from the Spey to eventually enclose Loch Spynie itself (Comber 1993). This sequence of events led to the emplacement of the extensive gravel ridges found presently in the vicinity of Lossiemouth (Ross 1992). At Culbin, the effect of the closure of the link to the Spey was to dramatically reduce the amount of sediment available for storm ridge sedimentation and set in motion a series of re-organisational phases as reflected in spit erosion, deposition and migration (Hansom 1999).

Holocene landforms

The sea level and sediment supply history has major implications for the geomorphology of the lower River Spey and Spey Bay. The Spey Bay area consists of a series of raised marine and fluvial deposits which underlie the entire area between Portgordon in the east to Lossiemouth in the west and in the lower Spey valley almost as far upstream as Fochabers (Figure 3.5). These deposits are backed by an extensive raised cliffline at ca. 9m OD. The cliff is 5-7m high, and is mainly eroded into Late Devensian glacifluvial and glaci-marine deposits (Firth 1989). In the Culbin area, Firth (1989) and Comber (1993) recognised all of the features seawards of the raised cliffline as Holocene in age, and this is also the case for Spey Bay.

Holocene raised gravel storm ridges are found at altitudes of up to 9.12m OD, against the foot of the glacifluvial deposits of the Holocene cliff. These gravel ridges are

discontinuously exposed and can be traced over 15 km from close to Porttannachy in the east to the sub-parallel ridges and recurving gravel arcs near Lossiemouth in the west (Figure 3.5). They represent abandoned upper beach deposits formed during storm events and are mainly composed of 40-50mm gravel. Initial emplacement of the ridges may have begun about the peak of the Holocene Transgression ca. 6500 BP, when rising RSL forced large quantities of material from the inner continental shelf onshore, which, together with Spey gravels, infilled the low and flooded areas south of the present coast.

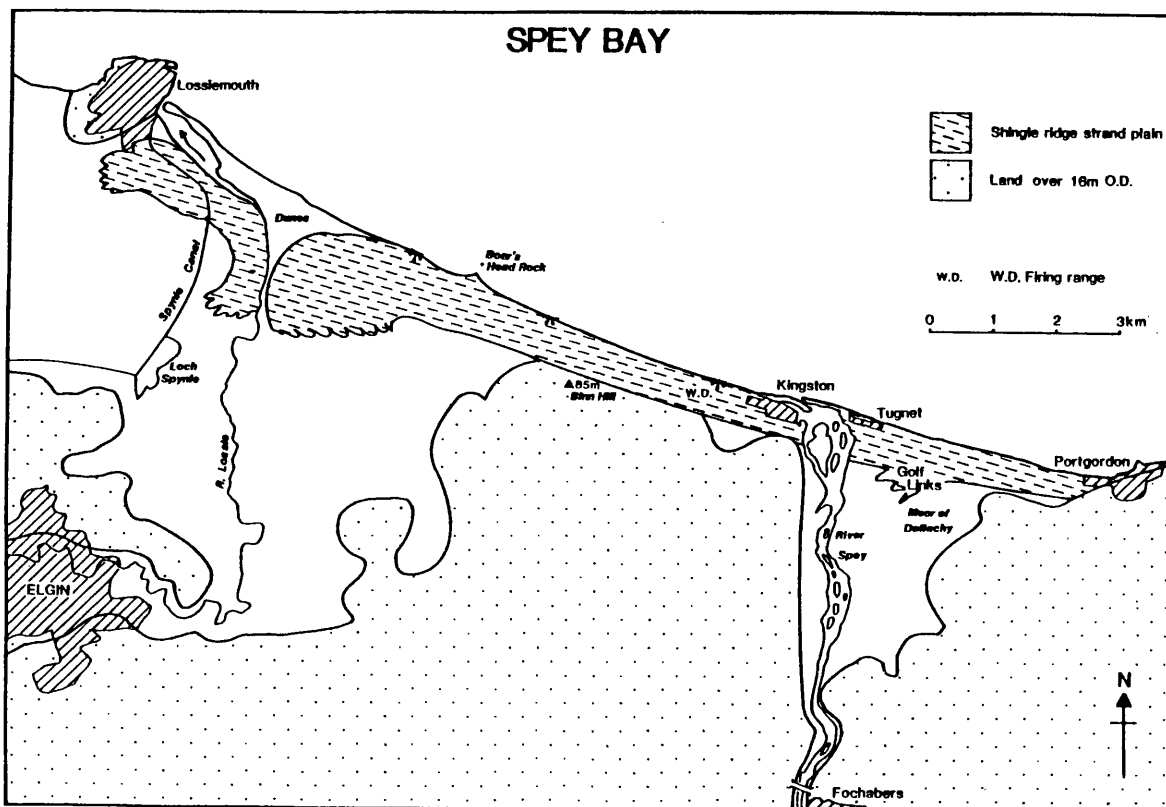


Figure 3.5: Spey Bay gravel strandplain (source: Ritchie 1983)

In the east the coastline extended over the area of the Moor of Dallachy and at least as far upriver as Warren Wood near Fochabers (Figure 3.5). In the west, the entire area of the lower Lossie/Spynie was an inlet of the sea so that Spey Bay was linked with Burghead Bay (Comber 1993). Progressively, the bay of the lower Spey was infilled by fluvial accretion behind gravel storm ridges developed across the mouth between Porttannachy and Kingston. Extending westwards from Kingston the shingle ridges fronted Binn Hill and accreted westwards to cut off the inlet of the River Lossie from the open sea. Such was the volume of gravel that eventually an 800m wide swathe of ridges developed to separate Binn Hill from the sea (Figure 3.5). The closure of the Lossie/Spynie Bay appears to have occurred as a direct result of longshore transport and reworking both of Binn Hill and Spey-derived gravels (Comber 1993). Some eastward-trending ridges and recurves extend into the former bay from the Lossie headland.

Altitudes of the ridges suggest that the majority of the sequence was deposited after the peak of the Holocene Transgression, with transects levelled across the sequence displaying a stable and then rapidly falling trend in altitude (Figure 3.6). The sequence of well-developed gravel ridges adjacent to the MOD firing range (centred on NJ315660) represent important marine indicators of a RSL above the present level and display ca. 1 m high ridges and troughs which fall from 9.12m OD to 2.44m OD where they merge with the rear ridges of the present coast (Figure 3.6).

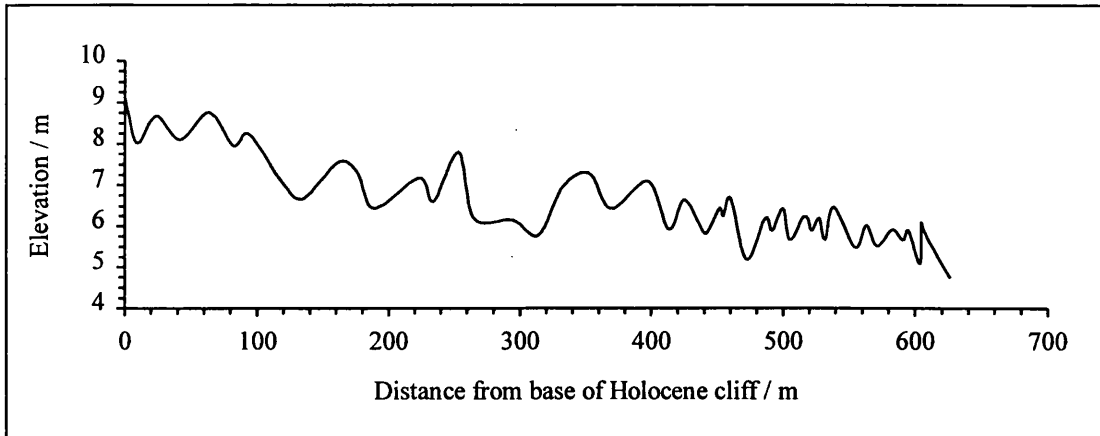


Figure 3.6: Transect of the raised gravel beach ridges at Spey Bay, from the base of the Holocene cliff to the present coast

3.3 The River Spey

The River Spey flows in a predominantly north easterly direction for a distance of ca. 157 km from Loch Spey to Spey Bay (Figure 3.7) draining a catchment area of 3 011 km² with a total stream network length of 36 400 km (NERPB 1995). The Spey drains the eastern slopes of the Monadhliath mountains and the northern slopes of the Cairngorms (Figure 3.7).

The Spey catchment is dominated by Palaeozoic metamorphic crystalline rocks and granitic intrusions (Maizels 1988) while the Moray Firth coastal plain is characterised largely by Old Red Sandstone conglomerate and sandstones (Section 3.1). The Spey river system was probably initiated in the mid-Tertiary, but numerous river capture events occurred during the Quaternary. Repeated glaciation and sequences of vertical movements in base level produced a distinctive combination of incised valleys and remnant plateau surfaces throughout the catchment (Maizels 1988).

The greatest impact of glaci-fluvial activity on the Spey valley was the accumulation of thick sand, gravel and boulder outwash deposits that extend across the valley floor and for tens of kilometres from Aviemore to Speymouth (Maizels 1988). These outwash deposits

were reworked into a prominent series of terraces by meltwater during progressive ice wastage. Upstream of Aviemore, distinctive high-level sand and gravel terraces up to 1.5 km wide occur on both sides of the Spey valley; in the Grantown area, a sequence of up to 5 terraces extends across the valley floor, with the highest lying over 30m above the level of the river (Brown 1871; Hinzman and Anderson 1915; Young 1977 cited in Maizels 1988). Glacifluvial and alluvial sand and gravel is widespread in the lower Spey valley from near Orton House (NJ313540) northwards to Speymouth. Most of this material occurs in terraces (Figure 3.8), but it is likely that the valley fill is several metres thick. South of Fochabers, the terraces are cut into pre-existing deposits of till and silt-sand glacial lacustrine sediments, with the terrace gravels lying on top in varying depths of up to 4m (Peacock et al. 1977). The vast volumes of sediment stored in the Spey terraces continue to provide a large potential supply of sediment to the coast.

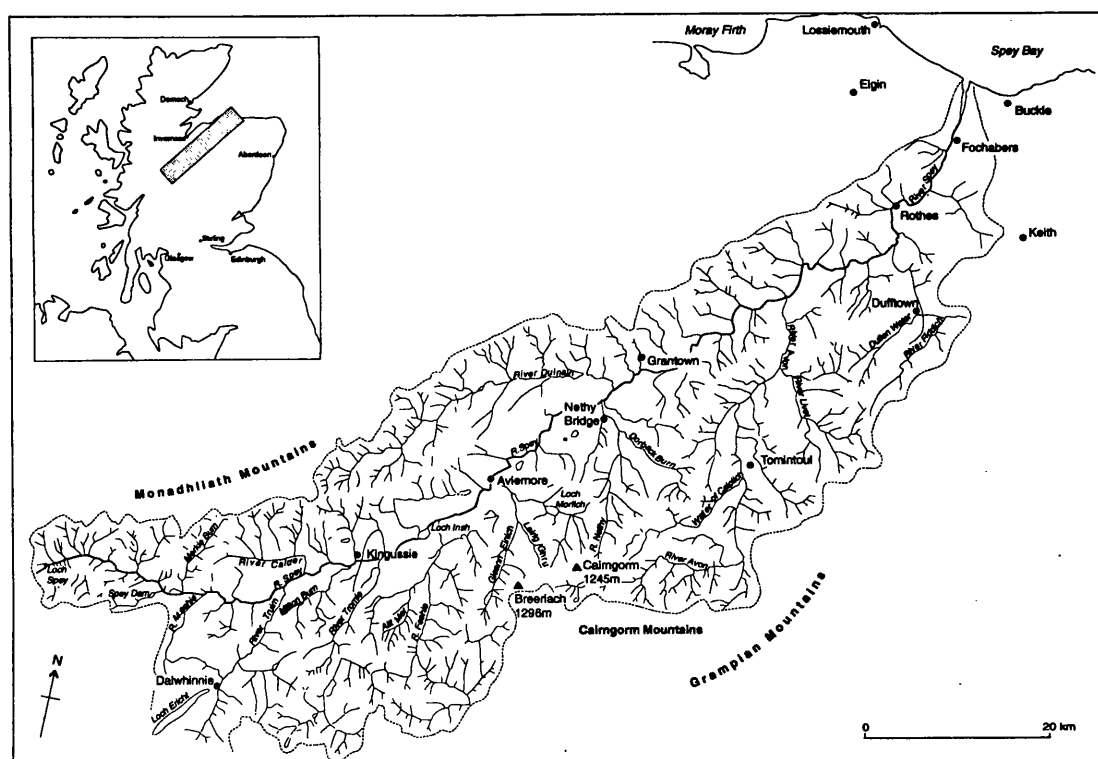


Figure 3.7: Drainage network of the River Spey (source: adapted from Maizels 1988)

The upper Spey catchment to Newtonmore is relatively steep (1:225) as is the lower river below Grantown-on-Spey (1:380) (Figure 3.9). However, the middle part of the catchment is characterised by a broad meandering channel and wide floodplain at a lower gradient (1:1200). The lower reach downstream of Orton to the coast maintains a steep gradient of 1:227 (4.4 m in 1 km) (Lewin and Weir 1977) and at the coast the gradient is 1:376 (2.66m in 1km), unusual for a large river so close to the mouth.

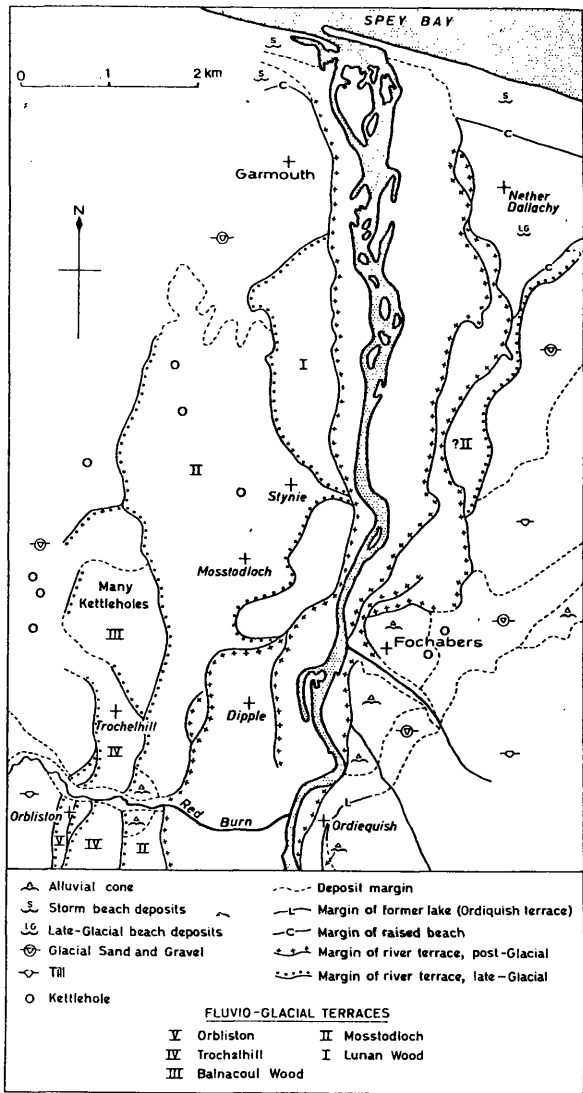


Figure 3.8: River terraces of lower Strathspey (source: Peacock et al. 1968)

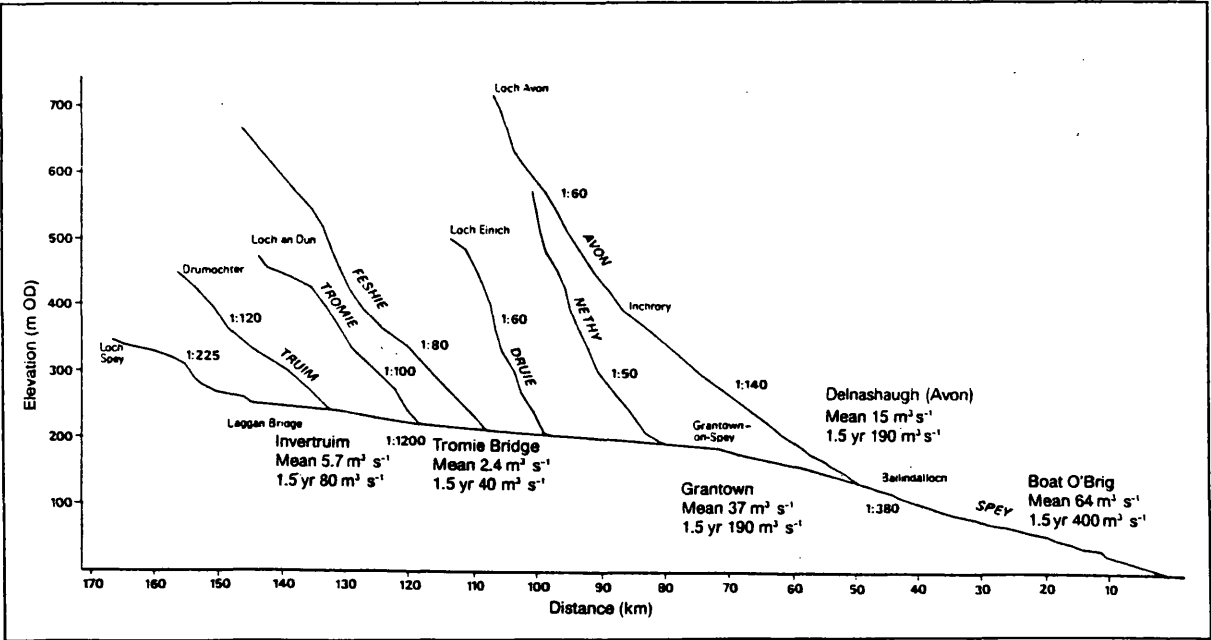


Figure 3.9: Longitudinal section of the Spey and its major tributaries (source: Inglis et al. 1988)

3.3.1 Catchment hydrology

The gauge at Boat o' Brig (NJ318518) is the furthest downstream and has a detailed flow record from 1952 to the present. The long-term mean flow is $64.4\text{ m}^3\text{ s}^{-1}$, with a range from $9.6\text{ m}^3\text{ s}^{-1}$ to $1675\text{ m}^3\text{ s}^{-1}$. The lowest flows generally occur during late summer, although annual minima have been recorded during extreme winter frost conditions. There is no general season for floods; the annual maximum floods at Boat o' Brig include spates in every month (NERPB 1995). The seasonality of flooding is complex and controlled by different mechanisms. Winter/early spring floods result from rain falling on snow creating melt in the upper Spey catchment whereas summer flooding is often generated in the Avon and other north facing tributaries as a result of frontal storms (Green 1958, 1971).

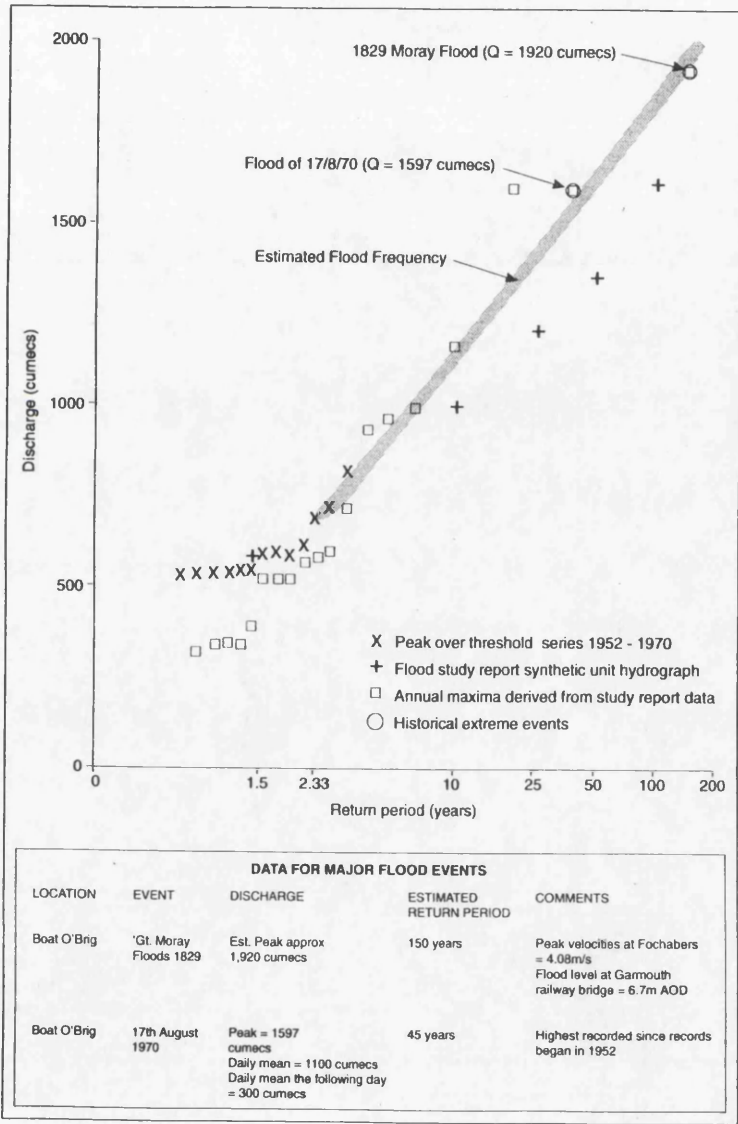


Figure 3.10: Flood frequency estimates for the River Spey at Boat o'Brig (adapted from Dobbie & Partners 1990)

Of the many notable floods of the Spey the most documented is the Great Moray Flood of 1829 (Lauder 1873; Wallace 1881), which had an estimated peak discharge at Boat o' Brig

of $1920 \text{ m}^3 \text{ s}^{-1}$ (Inglis et al. 1988). Long term flood frequency estimates show that the 1970 and 1829 floods have return periods of ca. 45 and 150 years, respectively (Figure 3.10). The mean annual flood ($Q_{2.33}$) and the median annual flood ($Q_{1.5}$) discharges for the lower River Spey are $695 \text{ m}^3 \text{ s}^{-1}$ and $485 \text{ m}^3 \text{ s}^{-1}$, respectively.

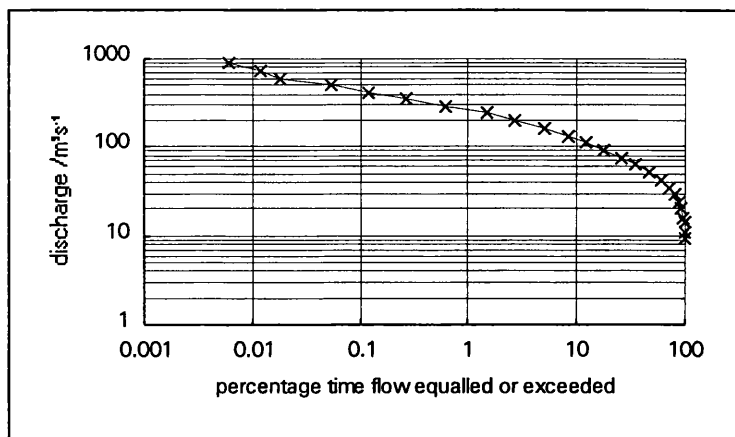


Figure 3.11: Flow Duration curve for the River Spey at Boat o'Brig (source: NERPB unpublished data 1953-1998)

Complete analysis of flow records recorded every 15 minutes at Boat o' Brig between 1953-1998 has been undertaken to produce a flow duration curve (Figure 3.11). It is the higher flows that are of interest in this study, as these cause the main morphological changes and induce high levels of sediment transport. Flows of $161 \text{ m}^3 \text{ s}^{-1}$ are equalled or exceeded 5% of the time (Figure 3.11).

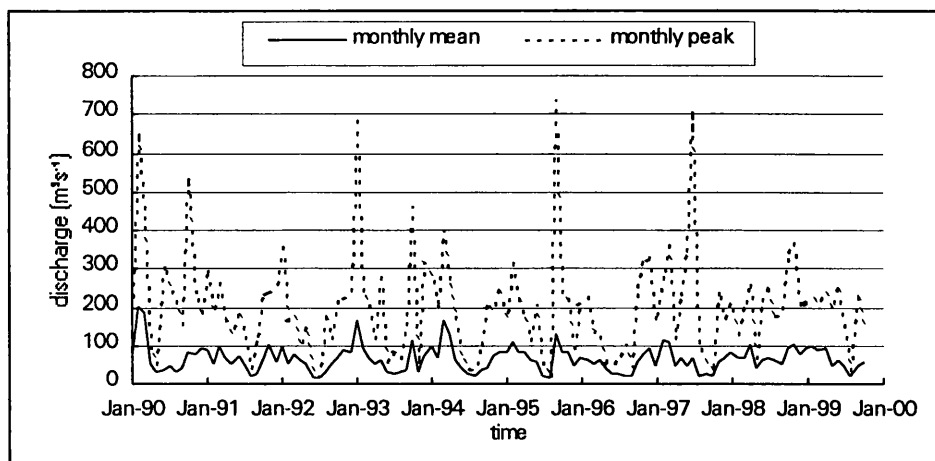


Figure 3.12: Monthly mean and peak flows of the River Spey at Boat o'Brig (1990-1999)

Flow records dating back to 1990 were analysed in detail. Six floods with peak discharges exceeding $400 \text{ m}^3 \text{ s}^{-1}$ occurred in the 10 year period (Figure 3.12) and the mean monthly discharge ranged from $200 \text{ m}^3 \text{ s}^{-1}$ (Jan-90) to $14 \text{ m}^3 \text{ s}^{-1}$ (Aug-95). In summary, the flow of the

lower River Spey is highly variable, experiencing a wide range of flows and a somewhat 'flashy' regime, with short periods of extremely high flow (Figure 3.13).

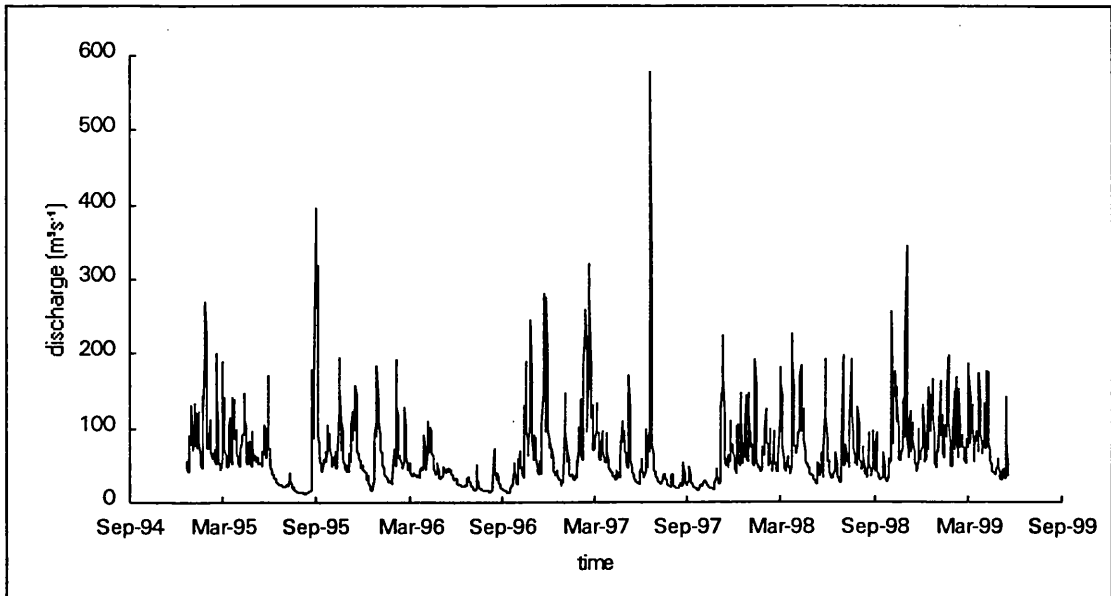


Figure 3.13: River Spey daily mean flows (at Boat o' Brig) from January 1995 to June 1999 (source: SEPA unpublished data).

River Flow during the field study period

Analysis of daily flows at Boat o' Brig during the study period (1995-1999) highlight two major flood events (Figure 3.13). The first occurred on the 10th of September 1995, with a mean daily discharge of ca. 400 m³s⁻¹ (Figure 3.13) and a flood peak of 730 m³s⁻¹ (Figure 3.12). This event occurred just before a period of field survey and caused major flooding and morphological change in the study reach. A flood of this magnitude has an estimated return period of ca. 3 years (Figure 3.10). The second major flood event occurred on the 1st of July 1997 with a daily mean of 577 m³s⁻¹ (Figure 3.13) and a peak of 705 m³s⁻¹ (Figure 3.12). Apart from these two events no floods with peak discharges greater than 400 m³s⁻¹ occurred.

3.3.2 General characteristics of the lower River Spey and floodplain

The lower River Spey is a high energy, wandering gravel-bed river. Toward its mouth there is a tendency for the river to become more braided, which may be a result of the steeper gradient (Section 3.3). Wandering gravel-bed rivers are relatively common world-wide (Section 2.2.1) but are rare within the United Kingdom (the only really comparable site is the River Feshie, itself a tributary of the Spey).

3.3.2.1 Geomorphology of the lower River Spey

The geomorphology of the lower River Spey and its floodplain (Figure 3.14, in sleeve) was mapped from 1994 aerial photographs and field survey undertaken in October-December 1995 (Section 4.1.3). The timing of the initial field mapping was coincidentally advantageous as it followed a major flood event in early September 1995 (Section 3.3.1). The occurrence of this event enabled the following additional information to be gained about the behaviour of the river in flood: locations of bank over-topping and directions of water flow over the floodplain; the extent of re-occupation of 'old' channels on the floodplain; identification of areas of the river which are sensitive to channel change (particularly bank erosion). The geomorphology is described below.

(a) The active river channel

Upstream of the viaduct: The ca. 2 km reach upstream of the viaduct has a series of meanders, some of which have cut-off channels along their inside bends, producing local braiding (Figure 3.14). Between these bends, the river is relatively straight, with occasional islands of bare gravel. In December 1995, the main flow turned through an angle of almost 90° at Essil bend (Figure 3.14). The channel configuration at this highly unusual bend has changed substantially during the study period and by October 1997 the western channel was entirely cut off, with the eastern channel conveying almost all the flow (see Section 5.2.3).

Viaduct to the coast: In this lower reach the river becomes more braided (Plate 3.1), with the individual braid channels continuing to meander. Downstream there is a tendency for the active channel width to increase (Figure 3.15). The active channel width is defined as the distance from bank to bank, excluding vegetated areas, whereas total channel width includes vegetated bars and islands within the limits of the active channel.

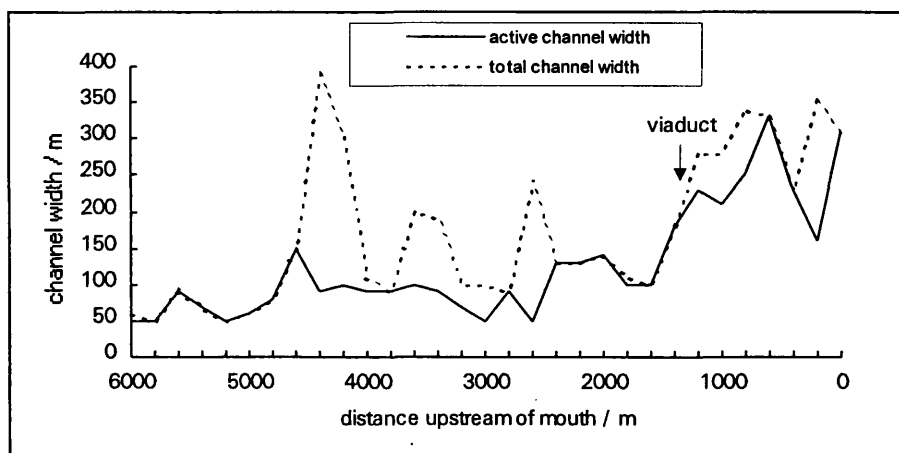


Figure 3.15: Variations in channel width of the lower River Spey (1995)



Plate 3.1: Oblique aerial photograph of the lower River Spey looking upstream from the mouth (July 1998). Note the increase in active channel width and braiding towards the mouth.

The increase in active channel width towards the coast reflects the depositional nature of the lower part of the reach. Variations in this parameter further upstream provide an indication of the locations of sediment storage within the active channel, a greater active channel width suggesting increased storage (cf. Church 1983). It is also apparent that the duration of storage varies, with the vegetated bars representing relatively long term storage and reduced bed activity (the age of the vegetation provides a minimum duration of storage for sediment in these features (cf. Nakamura and Kikuchi 1996)). The presence of vegetation on the bars does not necessarily indicate enhanced long term stability, and there are a number of locations where vegetated bars have recently been eroded (Plate 3.2).

(b) The floodplain

Prior to human activity, the lower River Spey floodplain was modified and re-worked as the active main channel migrated across its flood plain (effectively the area within the Holocene cliff limits, Figure 3.14). Many remnants of former main and minor channels remain and continue to convey Spey floodwaters during spates (although their frequencies of occupancy varies). Some are re-occupied annually, whereas others are inundated only during extreme flood events. During the September 1995 flood (Section 3.3.1) over 90% of the identifiable floodplain channels were occupied to water depths of up to 1.5m. The intensity of this flow can be gauged by the presence of erosional features such as scour of the channel beds and tree removal, and deposits such as gravel spreads within the channels.

3.3.2.2 Sediment characteristics of the lower River Spey

The sediment of the lower River Spey consists mainly of reworked glaci-fluvial sands and gravels (Section 3.3). Inglis et al. (1988) note that sediments are generally coarse gravel and cobble sized near the mouth, with a characteristic sediment size of 75 mm. Much larger material may be moved during floods and grain sizes larger than -7ϕ (128 mm) are commonly found (Lewin and Weir 1977).

The surface sediment samples collected in this study (see Section 4.4) are generally well sorted and symmetrical (Figure 3.16). The high degree of sorting of these samples ($\sigma_g = 1.7-1.9$) is indicative of this being a distal location, at distance from the sediment source. The material has thus been well sorted during transport and the well rounded bed material indicates efficient abrasion. There are variable amounts of sand present within the river bed at different locations downstream. Surface sand exposures are found in particular hydraulically controlled locations (for example, at the tails of bars) and grain sizes vary in response to local conditions.



Plate 3.2: Erosion of vegetated bar (NJ 345605), lower River Spey (October 1995)



Plate 3.4: Coastal erosion near Porttannachy (NJ 385644). Note cusps in the gravel and the eroding grassy bank landwards of the beach (October 1995).

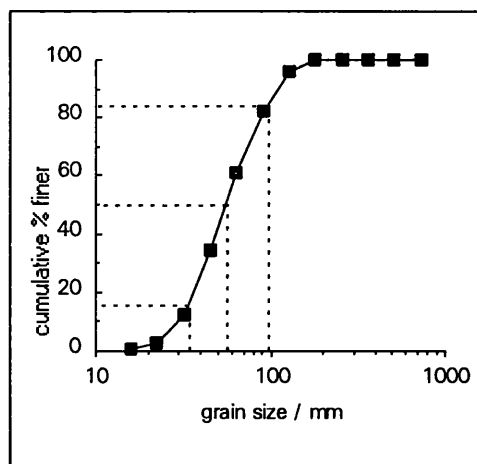


Figure 3.16: Typical surface sediment sample (Section 9) 3300 m upstream of the river mouth. This marks the upstream limit of the study site. $D_{50} = 55.57 \text{ mm } (-5.8\phi)$, $D_{84} = 95.10 \text{ mm } (-6.57\phi)$, $\sigma_g = (D_{84}/D_{16})^{0.5} = 1.67$.

There is no systematic downstream change in sediment size or sorting within this distal reach of the river (Figure 3.17), which has a mean D_{50} of 42 mm. However, there is an increase of about 50% in median (D_{50}) and D_{84} grain sizes between 2000 and 4000m upstream of the river mouth. This coincides with a positive deviation from the generally smooth long profile of the river in this area (Figure 3.17). Although the amount of data available is limited, this association indicates the possibility of a sediment storage zone in this location, which extends approximately 2km upstream from the railway viaduct (Gemmell et al. 2000).

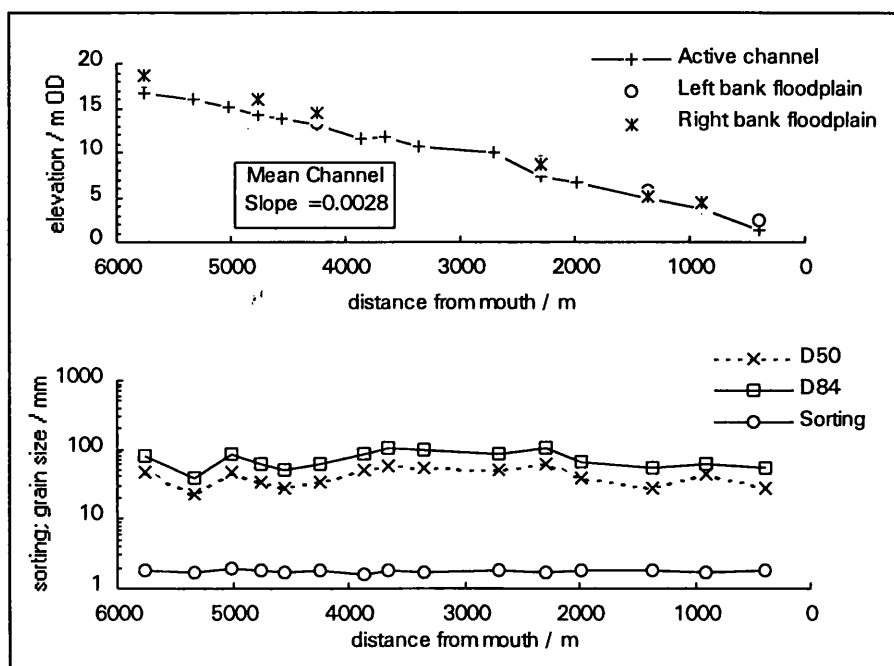


Figure 3.17: Channel elevation and surface grain size characteristics of the lower River Spey (surface samples collected at every profile location and based on a random sample of 100 clasts, see Section 4.4 for details). Average D_{50} is 42 mm.

Detailed analysis of bulk surface and sub-surface samples of fluvial sediment (Section 4.4) is presented in Chapter 5.

3.4 The Speymouth delta

The mouth of the Spey is amongst the most dynamic parts of the British coast (Plate 3.3, Grove 1955). Cartographic records analysed by Grove (1955), Ordnance Survey maps, aerial surveys and field mapping document the changes in the position of the river mouth between 1724 and 1995 (Figure 3.18 and Table 3.1).

Year	Comments	Source
1724	Note says mouth in same place as for 1860 westerly entrance, 1554m west of Tugnet.	2
1798	Speymouth spit 5km long (tradition although no map evidence)	3
1829	Great floods of Spey. Houses lost at Kingston and natural breach of spit	3,4
1844	Unsuccessful cut by Duke of Gordon, riparian proprietor	1,2
1857	Unsuccessful cut	1
1860	Successful cut brings mouth 232m west of Tugnet. Old mouth is 1554m west. (survey)	1,3
1867	Cut by the Duke of Gordon	2
1870	Mouth is 104m east of Tugnet, and takes an eastwardly course showing a reversal of normal tendencies. The old course is a lagoon (OS)	3
1885	Successful cut	1,2,3
1897	New cut made 232m west of Tugnet. Cut by Spey Fishery Board.	1,2,3
1903	Mouth is 232m west of Tugnet (survey)	3
1905	Mouth is 457m west of Tugnet. Lagoon has diminished since 1870. (OS)	3
1928	Note says mouth 674m west of Tugnet	3
1933	Cut made 232m west of Tugnet by Dept. of Agriculture.	1,2,3
1955	Mouth is 945m west of Tugnet near to Kingston and spit is well developed (OS)	3
1956	Major flood of the River Spey	3
1960	Major flood of the River Spey. Two houses in Kingston lost.	3
1960	Mouth is 1311m west of Tugnet, and spit is even better developed (aerial photography)	3
1962	Cut made by Moray County Council	1
1974/5	Cut made by Crown Estates	1
1981	Meandering river cut through spit unaided	1
1989	Cut made by Grampian Regional Council	1
1995	Growth of an ca. 400m long gravel spit, diverting the outlet west	

Table 3.1 : Historical record of the Speymouth spit and human efforts of realignment (sources: 1 GRC Roads Department Records; 2 Hamilton 1965; 3 Dobbie & Partners 1961; 4 Omand 1976).



Plate 3.3: Oblique aerial photograph of the Speymouth delta (July 1998) taken looking westwards from the village of Tugnet to Kingston in the far distance. Note the westerly trending spit diverting the mouth accordingly.

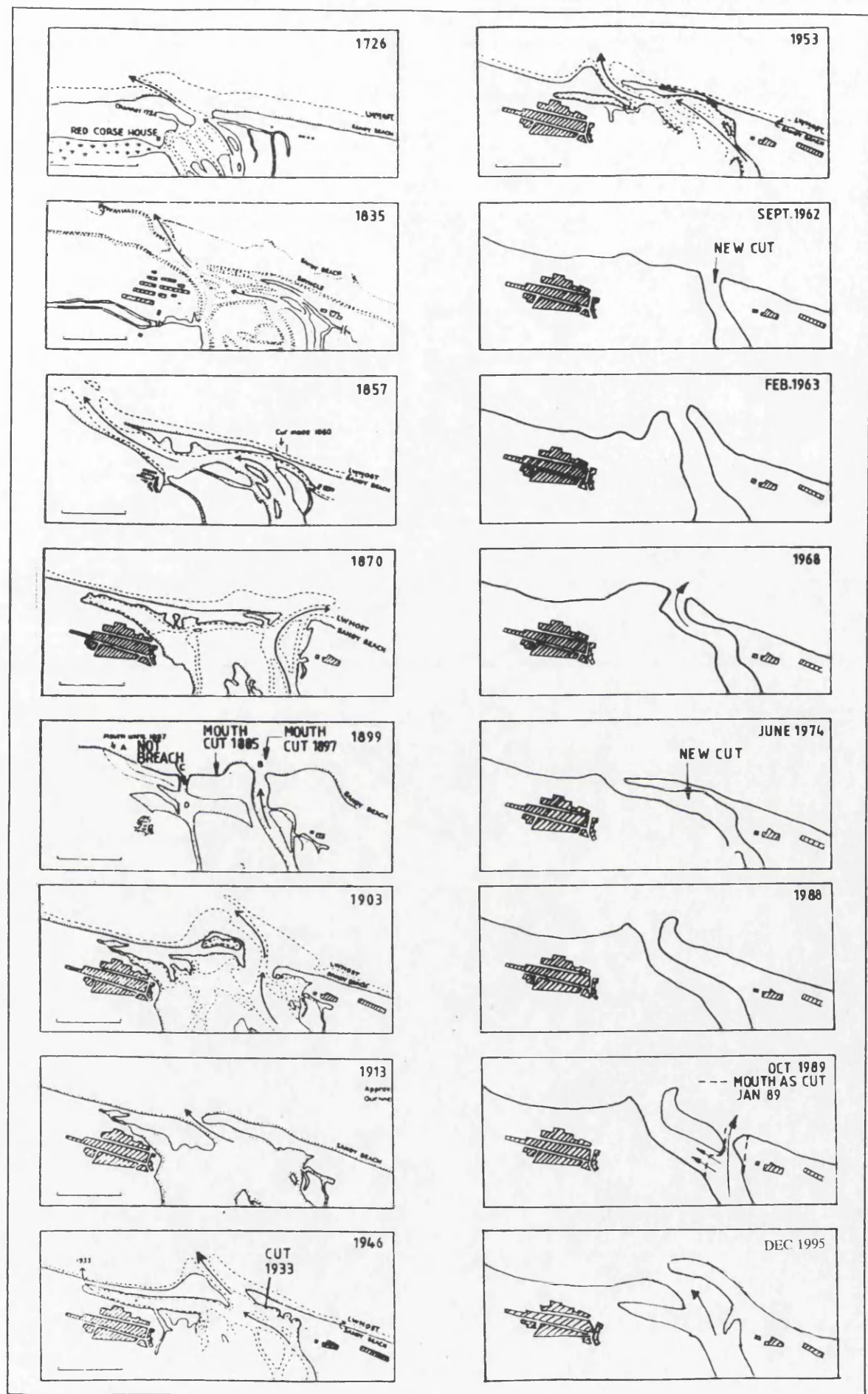


Figure 3.18: Evolution of the mouth of the Spey (1726-1995) (source: adapted from Dobbie & Partners 1990)

The natural tendency is for the river mouth to shift westwards towards Kingston, driven by the westward migration of the spit formations across the mouth (Table 3.1 and Figure 3.18). If this natural process is uninterrupted, historical records show that the mouth can migrate by as much as 1.2km west of its "central" position (ca. 200m west of Tugnet) and, according to local tradition, the Speymouth spit was ca. 5km long in 1798 (Hamilton 1965). The river has, in the past, returned to a central location through natural breaches of the gravel spit, as recorded in 1829 and 1981 (Riddell and Fuller 1995). Breaches have also been artificially engineered to realign the river mouth and reduce the threat of flooding and erosion at Kingston (Table 3.1). On two documented occasions in 1870 and 1989, the river outlet was diverted eastwards, showing a reversal of normal tendencies.

A complex suite of gravel ridges, enclosing tidally influenced lagoons, are present both to the east and west of the Spey outlet (Plate 3.3) and these can be used to locate former positions of the mouth of the Spey (e.g. the 1988 mouth was at the western tip of the lagoon and the 1963 river channel flowed through the lagoon north of Kingston). Gravel ridges relating to former river banks can also be identified. In July 1998, a gravel spit, prominent even at MHWS, extended ca. 400m westwards from Tugnet, diverting the Spey accordingly (Plate 3.3).

3.5 Spey Bay

3.5.1 Environmental conditions

The interplay of winds, waves, tides and currents shape the coastal geomorphology of Spey Bay, and are the driving forces of morphological change and sediment transfers at the coast. Typical environmental conditions at Spey Bay will be presented in this section, together with a record of coastal storm events during the study period (1995-1999).

3.5.1.1 *Wind climate*

The wind field along the south coast of the Moray Firth is dominated by south-westerlies channelled along the Great Glen. Analysis of a 10 year record of winds over 15 ms^{-1} recorded at RAF Kinloss, show a clear predominance in the sector 220° - 300° (Ross 1992). Wind data from Lossiemouth (1976-88) provides the most representative wind climate for Spey Bay (Figure 3.19). The wind rose shows a dominance of southerly and south-westerly winds (Figure 3.19). Only 35.4% of winds in the period 1976-88 come from the potential wave generating sector (Section 3.5.1.2), predominately from the north-west (300° - 360°), with a lesser occurrence of winds from the north-east (Figure 3.19).

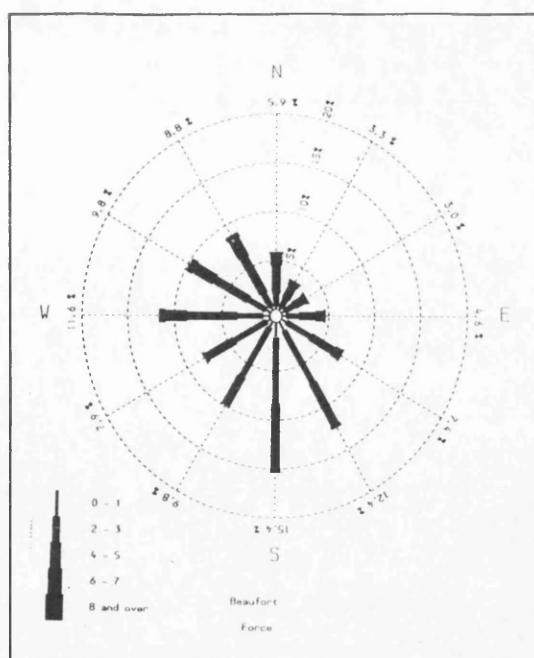


Figure 3.19: Spey Bay wind rose (1976-88) compiled from Lossiemouth wind data (source: Babbie Dobbie Ltd. 1994)

The maximum hourly wind speed at an elevation of 10m above Still Water Level (SWL) with an average recurrence of 50 years varies between 36 ms^{-1} at Fraserburgh to 38 ms^{-1} at Duncansby Head in Caithness (NERC 1992), with an assumed linear gradient at intermediate locations, so Spey Bay is ca. 37 ms^{-1} .

3.5.1.2 Wave climate

Wind waves are generated within the Moray Firth creating a broad spectrum of wave heights and periods. Swell waves are generated in the long fetches of the Atlantic and North Sea and travel into the firth, arriving at its coastline in a more ordered fashion. The entrance of swell waves from the North Sea is limited by the orientation of the Firth, creating an energy "window" in the sector 000° - 090° . British Maritime Technology (1986) suggest that incident waves from this sector occur for only 29% of the year. The remainder of the wave record is dominated by wind waves generated within the firth. Generation of wind waves in the vicinity of Spey Bay is limited by the orientation of Spey Bay to the sector 290° - 110° (Figure 3.19), although those from the north and east are not easily differentiated from swell waves. Monthly mean significant wave heights derived from Geosat Altimeter data for the outer Moray Firth (1986-1989) suggest that May is the stormiest, whereas February has the lowest waves (NERC 1992)(Table 3.2).

An offshore wave climate was derived for Spey Bay from wind data collected at Lossiemouth and Fraserburgh, using hindcasting (Dobbie & Partners 1990). Wave refraction analysis was undertaken, using the wave refraction model OUTRAY, to

establish inshore wave conditions at Spey Bay (Dobbie & Partners 1990). Probability distributions for waves during the period 1990-92 and over the previous 20 years (Figure 3.20) shows that the modal wave height occurring at Spey Bay is 0.4m, with heights of over 2m being relatively rare. Comparison of wave height and direction showed that the greatest number and largest ($> 4\text{m}$) of waves approach from the north-east. However, moderate waves arrive at the coast from all sectors.

Month	Monthly mean significant wave height / m
January	1.5 - 2.0
February	0.5 - 1.0
April	1.0 - 1.5
May	2.0 - 2.5
July	1.5 - 2.0
December	1.5 - 2.0

Table 3.2: Monthly mean significant wave height (highest 1/3 of all waves) for the outer Moray Firth (1986-1989) (NERC 1992)

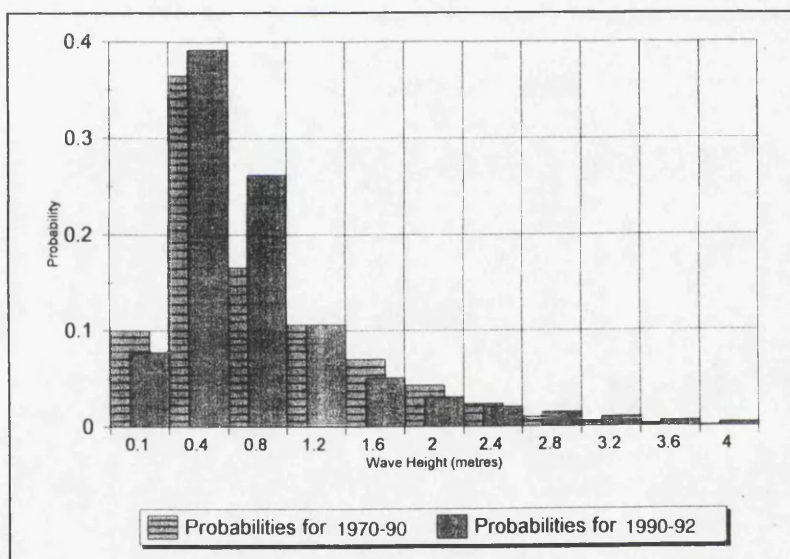


Figure 3.20: Wave probability distribution for the coastline between Kingston and Tugnet (source: Babbie Dobbie Ltd. 1994)

Extreme wave conditions with return periods of 10, 50 and 100 years were also predicted for offshore and inshore points in Spey Bay (Dobbie & Partners 1990)(Table 3.3). Extreme waves in the offshore zone are incident from the north-east sector (60°) as expected due to the greater fetch lengths from this sector. As the extreme waves move inshore to Speymouth they become attenuated and there is a small reduction in wave height to 5.79, 6.48 and 6.77m for return periods of 10, 50 and 100 years, respectively (Table 3.3).

Return period (yrs)	OFFSHORE			INSHORE		
	Wave Height (m)	Period (s)	Direction (°)	Wave Height (m)	Period (s)	Direction (°)
10	6.5	8.17	60	5.79	8.41	40.81
50	7.1	8.54	60	6.48	8.83	40.54
100	7.4	8.69	60	6.77	9.00	40.44

Table 3.3: Extreme wave conditions offshore and at the mouth of the River Spey (source: Dobbie & Partners 1990)

3.5.1.3 Tides

The Moray Firth is classified as mesotidal and experiences semi-diurnal tides with high water occurring approximately every 12.4 hours. The tidal range is limited and remains relatively constant along the outer coast from Lossiemouth to Buckie on the Spey Bay sector of the southern Moray Firth (Table 3.4). Spring tidal range at Buckie is 3.4 m, falling to 1.6 m on neaps.

	Spring Tidal Range (metres O.D.)	Neap Tidal Range (metres O.D.)
Lossiemouth	-1.5 → +2.0 (3.5)	-0.5 → +1.1 (1.6)
Buckie	-1.4 → +2.0 (3.4)	-0.5 → +1.1 (1.6)

Table 3.4: Tidal range at Lossiemouth and Buckie (derived from Admiralty 1993). Heights have been rectified to OD (Newlyn).

Whilst these values represent predicted tidal heights, values may vary considerably depending on meteorological conditions. For example a 10 mb drop in atmospheric pressure is capable of producing a 0.1m rise in the sea surface. Such storm surge conditions result in forced elevation of the sea surface, additional to wave set-up and enhanced water levels associated with cyclonic onshore gales. Predicted storm surge elevations with a return period of 50 years range from 1.25-1.50m in the vicinity of Spey Bay (NERC 1992). On-site observations suggest that the coincidence of a north-easterly gale and high spring tides can elevate water levels considerably along the coast, producing locally significant erosion (e.g. during the storm of March 1st 1998, Section 3.5.1.5).

3.5.1.4 Currents

Current flow patterns in the central Moray Firth basin are dominated by tidal flow, allied to channelling by fault-controlled orientation of sea bed features, closer inshore wave-driven longshore currents tend to prevail (Reid 1988).

Tidal Currents

NERC (1992) suggest that maximum tidal currents in the Moray Firth during mean spring tides are relatively low, ranging from 0.12ms^{-1} to 0.5ms^{-1} towards the inner reaches of the firth. Tidal currents were measured in the vicinity of Findhorn Bay, at a site 1.2km off Culbin Sands between June 5th - 22nd, 1991, spanning a neap-spring tidal cycle (Comber 1993). Recorded current speeds were generally low, with a modal peak of 0.05ms^{-1} and a maximum of 0.28ms^{-1} and the data suggested that the ebb tide was dominant. As some 43% of the record was below 0.05ms^{-1} and 84% below 0.10ms^{-1} (Comber 1993) they have little real effect on the transport of sediment. No data exists on tidal currents at Spey Bay, however there is no reason to expect them to be significantly different in magnitude from those measured at Culbin.

Wave Induced Currents

Wave induced currents are created when waves approach the shoreline, the most effective currents being found where waves impinge at an oblique angle. The refraction of wave crests as they approach the coast creates a division of incident energy contained within the incident wave train, with the resultant vectors of wave energy directed onshore and alongshore in fractions proportional to the angle of approach (Komar 1976; Leeder 1982; Pethick 1984).

At Spey Bay, the dominant angle of wave approach is from the north to north-east sector, due to the incidence of swell waves and greater fetch lengths from this direction (Section 3.5.1.2). Both the largest waves and the highest frequency of waves are recorded from this sector, although smaller wind waves generated in other sectors within the firth can be significant. The dominant energy influx from the north to north-east sector has strongly influenced the alignment of coastal features along the entire southern Moray Firth coast. The orientation of Spey Bay is 110° - 290° and, thus waves incident from the dominant wave directions meet the shoreline obliquely, producing a net westward drift of sediment.

Net westward drifting sediment at Spey Bay is manifest by the deflection of the spit at the Speymouth delta and the westerly extension of the active gravel beach ridge towards the sands at Lossiemouth (Section 6.1). Wave refraction analysis for the Spey Bay coastline

was used to model potential longshore sediment transport (Dobbie & Partners 1990). Results indicate that the potential westerly longshore sediment transport rate is ca. 3000 m³a⁻¹ (Riddell and Fuller 1995). However drift terms on a year-by-year basis were substantially different in magnitude and direction, demonstrating the sensitivity of the longshore drift system, and hence the beach, to the wave climate. There are several examples of short-lived easterly counter-drift at Spey Bay (Section 3.4).

3.5.1.5 Record of storm events (December 1995 - August 1999)

Table 3.5 documents the main storm events recorded at Spey Bay during the study period. There were three major storm events of note. The storm of late December 1995/early January 1996 caused ca. 5-10m of shoreline retreat at Kingston (Stratton, pers. comm.) and prompted the council to undertake beach recharge works (completed in March 1996, see Section 6.3). The next storm of note occurred in late December 1997/early January 1998, causing lowering of the beach crest at Kingston and significant over-washing of gravel into the lagoon. The combination of northerly gales, swell waves and high spring tides on the 1st March 1998 caused widespread damage along many parts of the Moray coast (Stratton, pers. comm.) and resulted in severe erosion along many parts of Spey Bay (see Chapter 6).

Date	event
late Dec 95 - early Jan 96	major storm (northerly gales)
Nov 96	northerly gales
Feb 97	severe gales (SW)
early March 97	severe gales (SW)
5/6 April 97	strong NW wind and high tide
4/5 May 97	very strong northerly gales
late Dec 97 - early Jan 98	major storm (northerly)
1/3/98	major storm (north-easterly swell and high spring tide)
27/28 May 98	strong north-easterly gale
3 March 99	very strong northerly gale

Table 3.5: Record of coastal storms at Spey Bay (December 1995 - August 1999) (sources: Moray Council, SNH records and field observations).

3.5.2 General characteristics of the beaches of Spey Bay

3.5.2.1 Geomorphology of Spey Bay

The present coastline of Spey Bay is marked by gravel ridges that are the seaward most representation of a raised gravel strandplain (Section 3.2) which separates the shoreline from a series of glacial and glaci-fluvial deposits and residual pockets of low lying marshy

ground (Ritchie 1983). The geomorphology was mapped along this ca. 16km stretch of coastline from the low tide limit to a landward limit defined by the Holocene cliff (Figure 3.21, in sleeve).

(a) Coarse-clastic beaches

The contemporary coarse-clastic beach extends over a distance of 13km and for convenience can be split into three areas; the beach east of the Speymouth delta to Porttannachy; the beach west of the delta; and the delta itself. The gravel landforms at the delta were previously discussed in Section 3.4.

East Spey Bay: The coastline between the low shore-platform cut in rock at Porttannachy and the well developed active gravel storm ridge near Tugnet has been described as a transitional coastline (Ritchie 1983) moving from a low mixed sand and gravel beach in the east to a ca. 4 m high gravel ridge near the mouth of the Spey. The active gravel ridge begins as a relatively low angle feature just west of the sea-wall at Porttannachy, where it has been obscured by rubble tipping. The ca. 10m wide gravel upper beach is fronted by a wide expanse of sand on the lower foreshore. Erosion is evident along this stretch of coastline, with recession of the upper shoreface and overwashing of gravel (Plate 3.4). Some areas of intertidal rock platform are exposed.

West of Tynet Burn the gravel ridge at the back of the beach becomes increasingly well defined and steeper and is often overtopped during storm conditions. This gravel ridge is continuous westwards and is fronted by a lower sand beach. Suites of well developed, but ephemeral, cusps of varying wavelength occur in the gravel beach face (Plate 3.5). The size and spacing of these features alters in response to short-term processes which vary with wave and tidal conditions.

At Norrie Scalp (between profiles +2 and +2.5km east of Tugnet (Figure 3.21) the gravel ridge at the back of the beach is narrower (less than 10m wide) and lower and fronted by a much wider expanse of intertidal sands and muds. Overtopping of the ridge crest is evident and large lobes of gravel have been deposited landwards of the beach onto the golf course behind. West of Norrie Scalp the gravel storm ridge regains definition, being some 20m wide with a crest height of ca. 5m OD. Along this stretch of coast, clear evidence of overtopping and recession of the gravel ridge exists and the coastal track, once used by the coast-guard, is now broken in several places.

A ca. 450m stretch of coastline at Tugnet has been protected by rip-rap placed at the back of the beach. Here the active storm ridge is well developed and the gravel beach is up to



Plate 3.5: Gravel storm ridge (east Spey Bay, NJ 379644), with a suite of well developed cusps in the beach face (March 1996)



Plate 3.6: Lossie sands and the westerly extent of the gravel beach in the far distance (NJ 261690). Note the wide intertidal sand beach backed to the landwards by high eroding dunes (May 1996).

80m wide towards the mouth of the Spey, with a number of landwards curving gravel ridges at the back of the beach. At the time of mapping (December 1995) the active gravel ridge continued as a defined ridge forming a spit across Speymouth, visible even at HWST (Figure 3.21). However, due to the dynamism at Speymouth this is a transient feature (Section 3.4).

West Spey Bay: The coastline to the west of Spey mouth contains the finest active gravel ridges in Scotland (Ritchie 1983), with steep gravel beach ridges extending ca. 9km west of the present mouth of the river before giving way to a low-angled sand beach backed by sand dunes some 3km west of Boar's Head Rock (Figure 3.21). There is substantial evidence that this gravel beach is accreting westwards encroaching on to the sand beach at Lossiemouth (see Section 6.1).

The width and height of the gravel beach varies considerably along the Bay, with the active ridge being generally lower close to Kingston and higher close to Boar's Head Rock. The beach ridges are widest to the west of Boar's Head Rock, where the beach is ca. 70m wide (Figure 3.21). The average altitude of the main gravel ridge crest is ca. 6m OD. Overtopping of the gravel ridge is apparent, particularly along the stretch of coastline from Kingston to the M.o.D. firing range. Here the active gravel ridge has undergone retreat, first burying, and then exposing on the foreshore, a line of World War 2 pill boxes and concrete tank traps, which originally followed the line of the back of the beach. The gravel beach is at it narrowest, around 15-20m wide, immediately west of Kingston.

Further west, close to the firing range, several low altitude vegetated ridge features are noted gently curving landwards and are truncated by the present gravel beach (Figure 3.21). These low altitude recurves occur continuously from this point to the westward extent of the gravel beach beyond Boar's Head Rock and become more defined and prominent westwards. West of Boar's Head Rock up to five gravel ridges curve gently landwards at the back of the present active ridge. At ca. 6m OD, these ridges stand up to 2m higher than adjacent intervening troughs. The landwards extremities of the ridges are sparsely vegetated with moss and grass before becoming obscured by sand dunes. A distinct break in slope followed by a 1-2m rise occurs landward of this series of recurves, which in many places marks the junction between the raised gravel strandplain and the more recent ridges. However, due to sand cover the break in slope cannot be identified further west. Truncation of the recurves by the present beach suggests that the last generation of gravel ridges was deposited along a coastline that trended along a west/east axis, rather than the present west-northwest/east-southeast axis and is strongly suggestive

of long term erosion and planimetric readjustment of this part of Spey Bay (Hansom and Black 1994).

(a) Sand beaches

Extensive sandy beaches are largely confined to the extremities of Spey Bay at Tannachy Sands in the east and Lossie Sands to the west (Figure 3.21). However, at the foot of the coarse-clastic beaches of Spey Bay, sand extends seawards as a low tide terrace, sometimes with patches of lag boulders, small rock outcrops and gravels.

Between Boar's Head Rock and the distal end of the Lossiemouth spit, the lower and middle sections of the intertidal area are of sand with bars which appear to be migrating westwards (Hansom and Black 1994). In the lee of Boar's Head Rock, the intertidal width increases to 150m on account of sand accumulation resulting from the disruption of longshore transport processes. The width of intertidal sand exposed increases westwards as the westerly accreting gravel tapers out, until the beach is entirely composed of sands backed by high dunes some 3km west of Boar's Head Rock (Plate 3.6).

The sand beach at Lossiemouth stretches for ca. 2.5km, fronting the dune and gravel ridge formations of the backshore and blending into low-angle intertidal sand flats. The beach is wide and relatively steep at its upper level, passing into a wide, gently sloping lower beach at approximately MHWS (Plate 3.6). An important element of the sand beach system is the transitional bare sand area forming the massive flat-topped bar that connects the main active aeolian deposits in the east with the dune remnants at the west end of the spit (Ritchie et al. 1978). This beach is between 100m and 300m wide and is characterised by the presence of stranded flotsam along its landward margin and topped by small embryo dunes (Figure 3.21). The highest parts of this wide expanse of open sand are infrequently inundated by the sea, but if storms occur during equinoctial spring tides, the recently formed embryo dunes undergo erosion and the sand is redistributed landwards across the back-beach and marsh surface.

(b) Sand dunes

Fully developed sand dunes occur only at the western end of Spey Bay close to Lossiemouth. However, along much of the shoreline from Porttannachy to Tugnet and from Boar's Head Rock to the end of the gravel spit, parts of the landward areas are thinly and intermittently veneered by a cover of blown sand. Many of the raised gravel ridges are also covered by this sand, particularly the gravel ridges at the junction with Binn Hill. This

suggests that in former times the downdrift beaches supported enough sand cover to allow the development of small dunes along their landward flanks.

The sand dune complex at Lossiemouth can be divided into two separate systems: the fully vegetated dunes at the proximal end of the Lossiemouth spit; and, the active mobile dunes at the distal end (Ritchie et al. 1978). The older, fully vegetated dune system at the proximal end is now eroding along its seaward flank (Plate 3.6). Migration of the gravel beach westwards (Section 6.1), although replacing sand by gravel, also appears to reduce frontal dune erosion by emplacing an energy-absorbing gravel ridge system at its toe. The dunes closest to Lossiemouth consist of a broad series of minor ridges which form a single, complex ridge up to 10m OD, broken by several overwash corridors (Figure 3.21). Frontal erosion is prevalent due to the proximity of the system to the seaward high tide mark, while redevelopment of the dune system by aeolian action is occurring vigorously in the lee hollows and behind the ridge. Extreme storm conditions further erode the dune system as waves spill over the upper beach through the gaps in the main ridge to drain south into the River Lossie. Sand frontally eroded from the exposed dune edge is soon redistributed by the wind to form new embryo dunes in the hollows, becoming quickly colonised by *Elymus* grasses (Ritchie et al. 1978).

3.5.2.2 Sediment characteristics of Spey Bay

Beach sediment

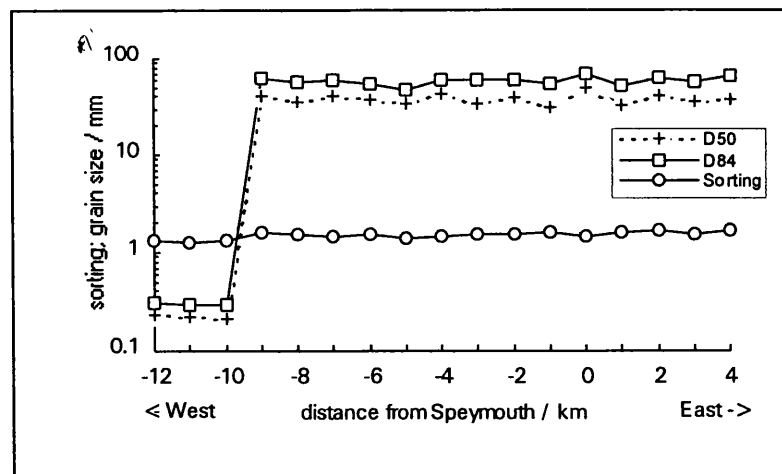


Figure 3.22: Grain size characteristics of Spey Bay. Sorting is $\sigma_g = (D_{84} / D_{16})^{0.5}$

The surface beach sediment, based on samples of 100 random clasts every kilometre along the coast, shows no obvious downdrift trends, until the abrupt transition from gravel to sand ca. 3 km from Lossiemouth (Figure 3.22). The median grain size (D_{50}) of the gravel varies from 30 mm to 50 mm along the beach, with a mean of 38 mm. The abrupt

transition to a sandy beach ($D_{50} = 0.22$ mm) occurs around 10 km west of the Spey mouth (Section 3.5.2.1). Detailed sediment analysis and results from bulk sampling are presented in Chapter 6.

Offshore sediments

Chesher and Lawson's (1983) map of offshore seabed sediments in the Moray Firth (Figure 3.23) show a large delta of coarse gravel around the mouth of the River Spey from the west of Kingston extending eastwards to Buckie. East of Buckie, sediments fine to sandy gravel with occasional gravelly sand, extending as far as Fraserburgh in a 10km wide belt (Chesher and Lawson 1983).

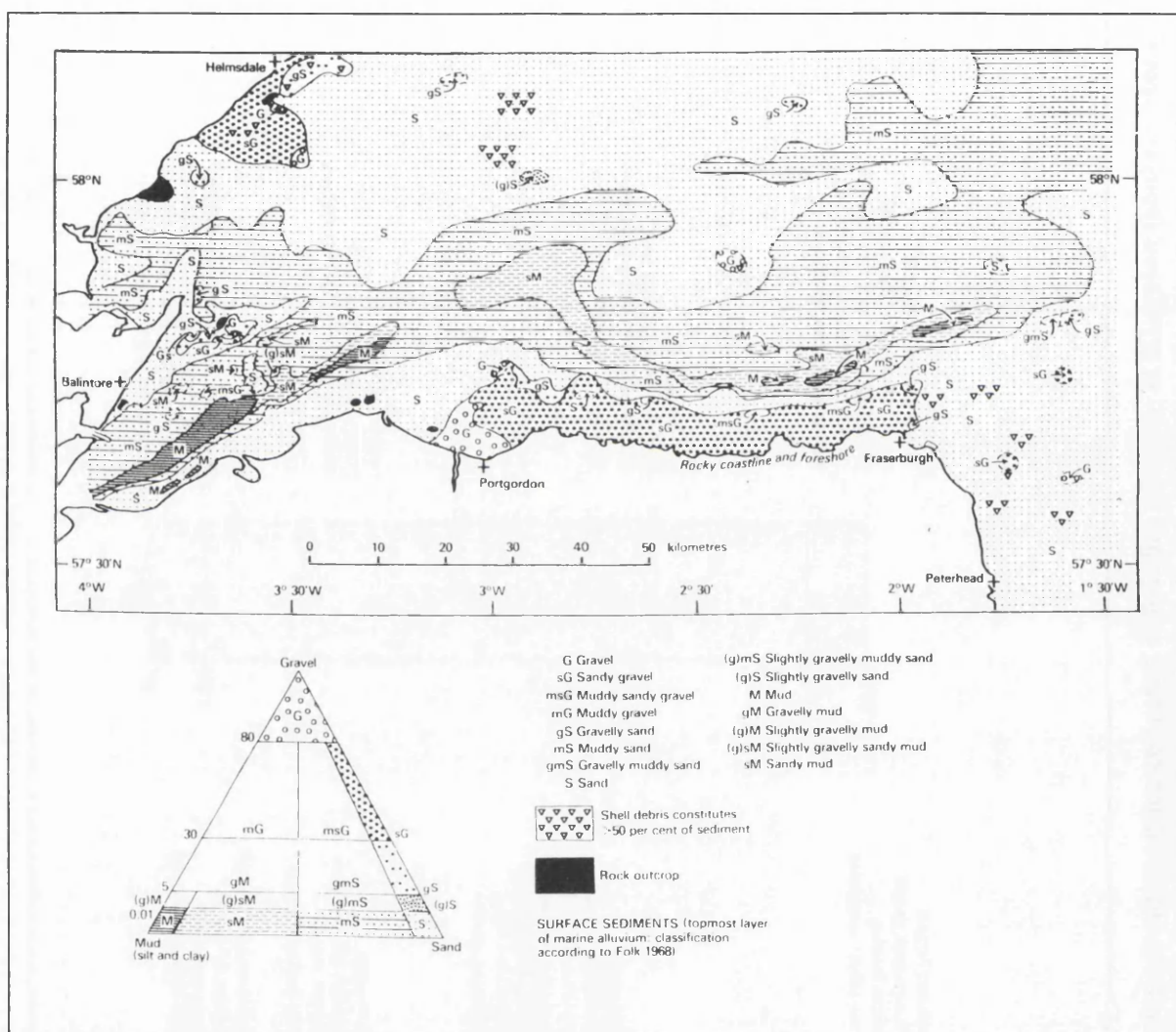


Figure 3.23: Seabed surface sediments of the Moray Firth (source: Chesher & Lawson 1983)

The apex of the submarine delta at the mouth of the Spey lies ca. 8.2km offshore, with a landward base ca. 10.9km wide, producing a surface area of 44.7 km² (Figure 3.23). A borehole at the seaward extremity of the fan (BGS borehole 71/15, Andrews et al. 1990, Section 3.1) revealed two distinct gravel units. The upper gravel unit was 3m thick,

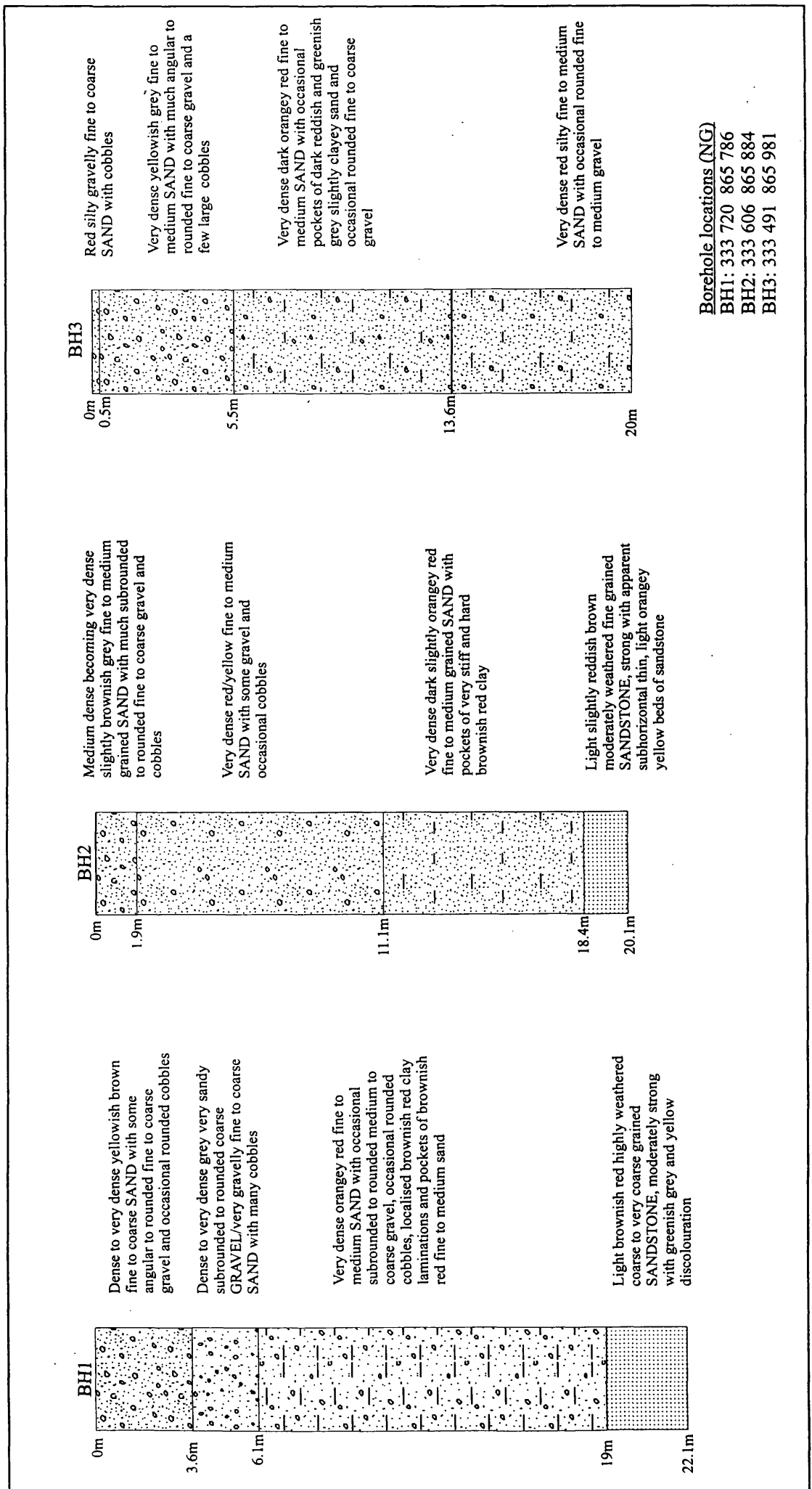


Figure 3.24: Borehole logs from samples collected 100-200m offshore at Kingston (source: Moray Council 1999 unpublished data)

separated from a lower, 2m thick, unit by 16 m of gravelly sand. The volume of gravel within this fan was calculated to be $2.24 \times 10^8 \text{ m}^3$ (Comber 1993), although this represents an underestimate as gravel present in the "gravelly sand" unit was not accounted for. The existence of substantial offshore sands and gravels is supported by the presence of a licensed off-shore dredging area (approx. $9.5 \times 4 \text{ km}$) which lies 4.6km offshore.

Babtie Dobbie Ltd. (1994) collected nearshore surface sediment samples at six locations between 0.5-1.5 km off the coast at Kingston and Tugnet to confirm the extent and sediment characteristics of the submarine delta. This sampling and analysis indicated that the nearshore sea-bed surface comprised of coarse to fine sands (Babtie Dobbie Ltd. 1994).

Three 20m deep boreholes were sunk ca. 100-200m offshore from Kingston. Results indicate that the sea bed sediment is predominantly sand with occasional rounded medium to coarse gravels in the deposit (Figure 3.24). A weathered sandstone bedrock was reached at depths of ca. 19m.

3.6 Summary

The active, wandering gravel-bed reach of the lower River Spey experiences a vast range of flows characterised by short duration flood events. The constantly changing morphology and channel pattern of the lower Spey make this an ideal site to apply a morphological approach to estimate sediment transport rates. Delivery of sediment to the coast occurs via a highly dynamic and constantly changing delta complex (Figure 3.18). The gravel beaches of Spey Bay are some of the most dynamic in the UK and are exposed to high energy wave and storm conditions, providing an opportunity to estimate gravel transport rates in response to storms of varying frequency and magnitude. Thus, the Spey system was chosen for this study as gravel sediment is transferred from the fluvial system to the coast in a highly dynamic environment.

4. DATA COLLECTION AND ANALYSIS

The sediment budget approach to quantifying sediment transfers was introduced in Chapter 2. This chapter outlines the application of this approach to the estimation of sediment transfers from the lower River Spey to the beaches of Spey Bay. The large, wandering gravel-bed river, the constantly changing morphology at the delta and the highly active gravel beaches of the field site (Chapter 3) necessitate some modifications to the methodologies described in Chapter 2, most of which were developed on much smaller scale systems. The methods of data collection are described and issues concerning the appropriate temporal and spatial scale of re-survey are discussed in this chapter. It is recognised that whilst detailed data collection is required to accurately quantify sediment transfers, this has to be balanced with logistical constraints. Error analysis of the required profile/survey point spacing to estimate sediment storage volumes and volume changes to within an acceptable error margin is also discussed.

4.1 Quantification of fluvial sediment storage and transfers

Contemporary fluvial sediment storage and transfers were estimated using a combination of three main techniques: repeat surveys of a series of cross-sections; repeat surveys of sediment storage elements (i.e. bars and islands) and; comparison of planimetric changes by repeat geomorphological mapping.

Reach scale surveys of sediment storage features in such a large river are time-consuming. Detailed surveys were carried out with two main objectives in mind. Firstly, to quantify the volume of sediment in *storage* in the fluvial system (e.g. Kelsey et al. 1987) and, secondly, to quantify the sediment volume *changes* over time within a reach, for use in estimating sediment transfers (Section 2.1.1.2). Throughout the three year study period there was only one morphologically significant flood event (July 1997, Section 3.3.1). This event induced significant channel change, particularly downstream of the viaduct (see Figure 5.9) and occurred prior to completion of the first detailed reach survey. Following the event the reach was re-surveyed and volumetric changes calculated for those sections of the river that had previously been surveyed.

Between July 1997 and December 1999 no significant flood events occurred (Section 3.3.1) and the bare gravel bars have become more stable and, in places, support dense vegetation. The channel pattern has become less complex and has stabilised into a more meandering configuration (especially downstream of the viaduct). The main morphological changes are significant bank erosion on the outer meander bends and deposition on the

inner bends (see Figures 5.9 and 5.10). As the bar sediment has been predominantly in temporary storage (at least over the period of study) detailed surveys of the sediment storage features (Section 4.1.2) will not measure any *change* in sediment storage over time. However, they can be used to estimate the volume of sediment in storage, which represents a potential sediment supply to the coast. Bar elevations have also been used to convert areal changes measured from repeat planimetric mapping (Section 4.1.3) into volumetric changes. Therefore, quantification of the fluvial sediment budget over the three years relies on a combination of cross-section surveys and repeat mapping. The limitations of this approach are discussed in Chapter 7. Fluvial data availability is shown in Appendix A.

4.1.1 River cross-sections

Suitable cross-section locations were identified from the base geomorphological map (Figure 3.14) to include large sediment storage features, such as mid-channel bars. In this, cross-section spacing is critical (Section 2.1.1.2) as changes in cross-section area at each section are assumed to be representative of the sub-reach between half distances to adjacent upstream and downstream sections. A section spacing of ca. 2-3 channel widths was chosen and sections were located at representative sites (Figure 4.1). A Wild T1010 Total Station (accuracy $\pm 0.005\text{m}$) was used to produce a network of 11 monumented cross-sections with an average downstream spacing of ca. 325m, linked at both upstream and downstream ends to OD (Ordnance Datum) via OS Bench marks. Sub-aqueous sections were surveyed from an inflatable boat, which was pulled across the channel via a locating rope, anchored to each bank. Depths less than 2.5m were determined directly using the survey prism; where depths exceeded 2.5m, the water surface was surveyed and the depth determined using a weighted plumbline. Cross-sections were surveyed on 3 occasions. The first survey in Dec 95/ Jan 96 did not include the wetted channel, whereas surveys in October 97 and May 99 did.

4.1.2 Sediment storage features

Detailed surveys of the river bars and islands were also carried out using a total station. Error analysis to determine an appropriate survey density is described in section 4.6.1. This analysis suggested a survey point spacing of ca. 4m.

4.1.3 Geomorphological mapping

Geomorphological mapping was initially carried out during a five week period between October and December 1995 using 1:5 000 aerial photography and ground checking. Maps were drawn at this scale and then photographically reduced to 1:10 000 for presentation.

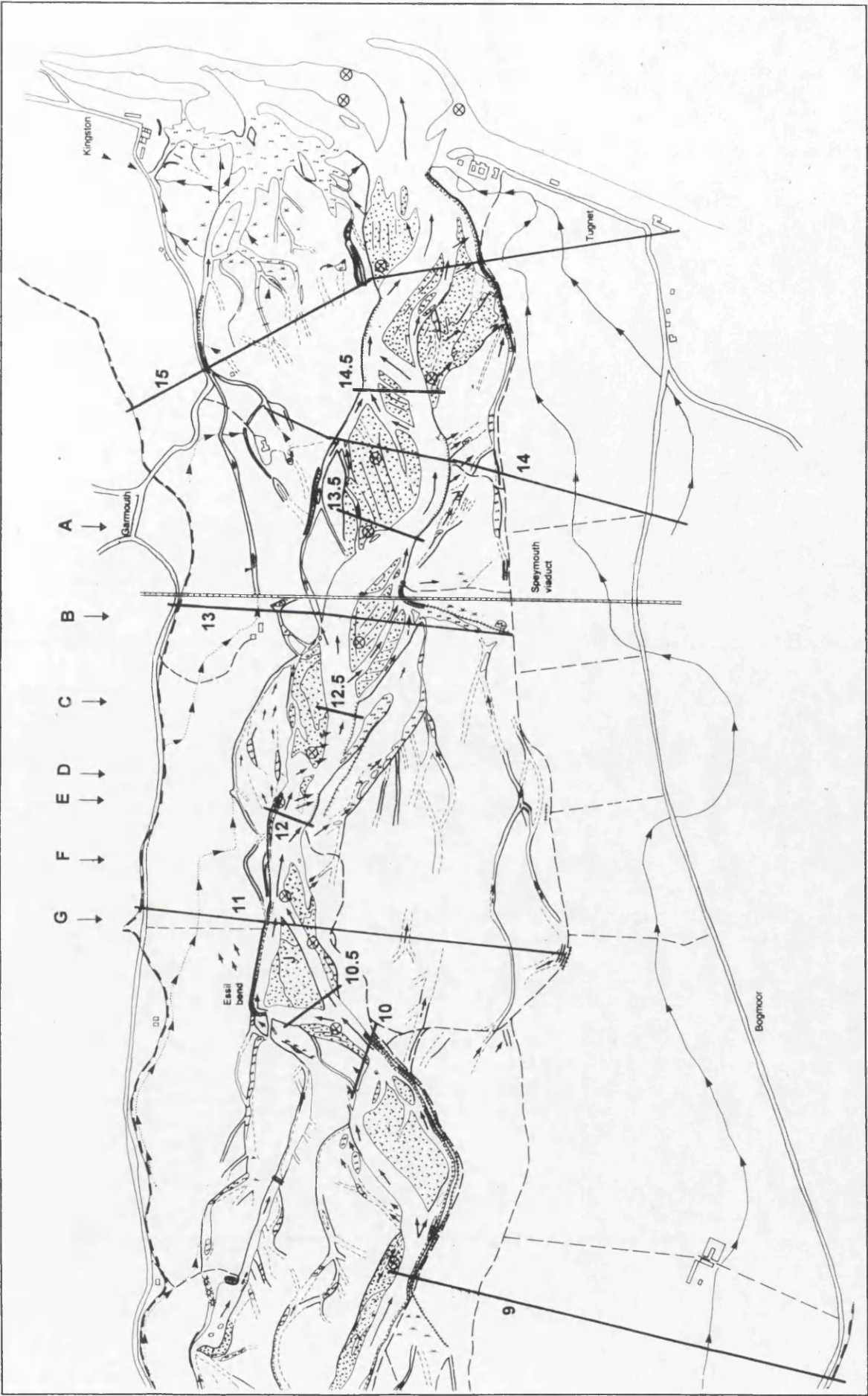


Figure 4.1: Cross-section and sediment sample locations on the lower River Spey. ⊗ marks sample locations.

Due to channel change throughout the study period (Section 5.2) the fluvial geomorphology map was updated in June 1997, December 1997 and April 1999 (see Figures 5.9 to 5.10).

To quantify planimetric changes in the lower Spey, the four geomorphology maps were digitised in CorelDraw with 14 Ordnance Survey control points on each map. The digitised maps were overlain in the GIS package Arcview, using the Ordnance Survey points to ensure accurate geo-referencing. For each period, polygons of change (i.e. erosion or deposition) were created in the GIS and their areas quantified. Areas were converted into volumes by multiplying by the depth of mobile sediment (d) obtained from bar (Section 4.1.2) and cross-section surveys (Section 4.1.1). To estimate volumetric changes prior to commencement of the field study, the main channel in July 1993 and July 1994 was digitised from aerial photography (at scales of 1: 10 000 and 1: 5 000, respectively) and input into the GIS for similar analysis.

4.2 Quantification of sediment storage and transfers at the delta

Repeat surveys of the sub-aerial and sub-aqueous delta morphology were used to quantify sediment storage and transfers at Speymouth. In addition, observations and photographs taken during beach profile surveys (Section 4.3.1) qualitatively document the constantly changing deltaic morphology of ridges, spits and channels.

4.2.1 Subaerial delta morphology

Subaerial sediment storage and transfers were estimated using a similar method to that used for the fluvial system (Section 4.1.2). Detailed surveys of delta morphology used a survey point spacing of ca. 4m (see Section 4.6.1 for justification). The coarse-clastic subaerial delta morphology was surveyed to a landward limit delimited either by the lagoon, salt marsh or rip-rap (Figure 3.21) and a seaward limit of low water mark (LWM). The volume of sediment stored at the delta above LWMS (-1.4m OD) can be calculated from this survey data. The subaerial morphology was surveyed in May 1997 and again in May 1998.

4.2.2 Sub-aqueous delta morphology

The sub-marine part of the delta required a detailed survey of the nearshore bathymetry. This survey was carried out on two occasions (August 1998 and August 1999) in order to assess any morphological changes. However, due to a large swell (ca. 1-2m) during the first survey it is considered less accurate, but serves to provide a general impression of the nearshore bathymetry.

Delta bathymetry was surveyed from a small fishing vessel using a Sonarlite echo-sounder for depths. The 1998 survey used a Geotronics GPS for XY positions and water surface elevations were determined using TIDECALC version 1.1 software (UK Hydrographic Office 1994) to predict water levels every 10 minutes through the tidal cycle. The 1999 survey used a Leica 500 Global Positioning System (GPS) for XYZ surface position. Depth was recorded every second as the boat moved slowly along delta transects, which converged on the river mouth. The GPS was operated in differential mode with a reference station on land and a rover station on the boat. Real-time XYZ co-ordinates were recorded every 10 seconds and boat elevations (Z co-ordinates) were converted to water level elevations using an appropriate offset. The GPS data are accurate to within $\pm 0.05\text{m}$ and depths to $\pm 0.05\text{m}$.

4.3 Quantification of coastal sediment storage and transfers

Coastal sediment storage and transfers were quantified via repeat surveys of a network of beach cross profiles (Figure 3.21). As it is critical that the change in beach profile area is representative of the changes along the section of coast it represents (Section 2.1.2), a detailed pilot study was carried out to assess the beach profile spacing required to quantify volumetric changes to within an acceptable error margin (Section 4.6.2). Clearly, a balance must be made between the level of accuracy and logistical constraints. For example, a profile spacing of 10m provides accurate estimates of volumetric changes but would require 1600 profiles and a field effort of some 150 days!.

4.3.1 Beach profiles

Thirty seven beach cross profiles were surveyed using a Wild Autoset Level (accuracy $\pm 0.005\text{m}$). Each profile (every ca. 500 m along the coast) was marked by 1m long survey pegs located on a stable surface landward of the main beach ridge. Repeat surveys were taken on a fixed bearing from the peg. A Wild T1010 Total Station was used to link the beach survey to OD, via OS Bench Marks at Porttannachy (NJ392642), Tugnet (NJ349653) and Gladhill farm (NJ323652). Shore-normal beach profiles were surveyed on nine occasions with an average temporal spacing of ca. four months. The timing of survey varied in response to spring tides and storm events (Appendix A).

These beach profiles were extended to ca. 500m offshore in August 1998 and August 1999 via nearshore echo-sounding. As an additional position fix, the boat was kept on the profile bearing using two aligned, highly visible, targets on the beach. Survey procedure and equipment is the same as in the delta bathymetry survey and is described in Section 4.2.2.

Although the 1998 survey is considered less accurate due to the large swell conditions at this time, it provides some data to assess nearshore morphological change.

4.3.2 Geomorphological mapping

The coastal geomorphological mapping was carried out between October and December 1995 using the method outlined in Section 4.1.3. The map includes the active beach, extends landwards to the Holocene cliff (Figure 3.21) and was updated using observations made during beach profile surveys.

4.4 Grain size measurements

The characterisation of the particle size distribution of coarse-clastic size sediments is problematic (e.g. Church et al. 1987; Gale and Hoare 1992; Ferguson and Paola 1997) for two reasons. Firstly, the range in grain sizes is often so wide that it becomes impractical to maintain a single method of measurement. Secondly, the degree of both lateral and vertical sorting and the structural features within the sediment bodies (Sections 2.2.1 and 2.2.3) can create bias in the sample size distribution depending on where the sample is collected (for example the sampled sediment size distribution of a coarse-clastic beach will vary between a cusp horn and a cusp bay).

There are two main methods of obtaining a representation of the sediment size distribution of coarse-clastic sediments both of which are used herein:

1. Grid-by-number sampling (Wolman 1954) is commonly used to characterise the surface texture of fluvial sediment, and is recommended by Kellerhals and Bray (1971) as the results are directly equivalent to bulk sieve analysis of subsurface sediments (Rice and Church 1996). It is important that the sample size generates a distribution that is statistically significant. Wolman (1954) recommended a 100-clast sample for a statistically significant estimate of the median grain size (D_{50}). More recently, Rice and Church (1996) recommended a sample size of 400 to obtain statistically significant estimates of percentiles within the grain size distribution (i.e. the D_{95}). A 400-clast sample gives an estimate of each percentile with 95% confidence limit of approximately $\pm 0.1\phi$ and any improvement in precision is achieved only at the expense of a much greater sampling effort (Rice and Church 1996).
2. Bulk sediment sampling generally involves sieving a shovelled or scooped bulk sample and apportioning each size grade by weight (Church et al. 1987). The sample provides a representation of the entire particle-size distribution of a sediment body. Accurate estimation of this distribution is critical for sediment budget studies in order to quantify

the size distribution and proportions of gravel and sand contained in any volumetric change. As coarse-clastic sediments (both fluvial and coastal) contain a wide range of sizes, determining the mass required to obtain a representative sample of each size fraction present in the deposit is problematic (Gale and Hoare 1992). Church et al. (1987) demonstrate that ≥ 100 particles are required in each 0.5ϕ fraction in order to obtain a stable measure of the proportion of that fraction in the entire sample. They recommend that the largest particle in the coarsest stable size fraction should constitute no more than 0.1% of the total sample mass (Church et al. 1987). This value is raised to 1% for maximum particle sizes of 32-128mm (Church et al. 1987). This limit is the target for all bulk samples in this study.

As beach sediments are generally better sorted than fluvial sediments this provides a conservative (i.e. large) sample size for beach bulk samples. Rigidly adhering to Church et al. (1987) criteria yields sample sizes of over 1000kg (Section 5.3.2). In order to speed up the field procedure, whilst maintaining a statistically significant representation of the grain size distribution, samples were truncated at 64mm. Everything greater than 64mm was classified in the field using a grain size template and weighed. The entire mass of sediment less than 64mm was weighed and then a sub-sample sieved and weighed. The sub-sample mass was determined based on Church et al. (1987) criteria as follows. Assuming a density of 2650 kgm^{-3} (i.e. quartz), a 64mm clast weighs ca. 0.36kg. Thus a sub-sample mass of at least 36kg of all sediment less than 64mm was sieved in the field. This was again truncated at 8mm and a sub-sample of ca. 400g was taken to the laboratory, dried, split and sieved.

The four main objectives of grain size sampling in this study are: (a) to describe any general downstream or downdrift trends in surface sediment size; (b) to assess cross-shore trends in beach sediment size, as morphological changes at high water and low water level may involve different calibres of sediment being transferred; (c) to obtain a statistically significant measure of the entire particle size distributions of the river, delta and coastal sediment for use in sediment budgets and; (d) to ascertain the characteristics of the offshore sediment in order to assess sediment transfers that may occur between the subaerial beach and the nearshore.

To achieve objective (a), beach and river sediment was sampled at every second profile location (Figure 4.1). Large grain sizes (8-180 mm) were classified in the field using a grain size template and a random sample of 100 surface clasts was taken (as recommended by Wolman 1954) using the pacing technique (Church et al. 1987). If finer sediment (< 8

mm) was present a sample of ca. 400g was taken to the laboratory, dried, split and sieved. Beach sediment was sampled at HWM. River sediment was sampled at the waters edge of the nearest bar to the profile and any fine sediment on the bar was sampled and sieved in the laboratory (Figure 4.1). In order to satisfy the criteria of Rice and Church (1996), six random samples of 400 surface clasts were also taken from bar-head locations (Figure 4.1).

To characterise any trend in cross-shore sediment sorting (objective (b)) 100 surface clasts were randomly sampled at HWM, mid beach and LWM at 10 profile locations. In addition, qualitative information of cross-shore sorting was obtained by taking vertical photographs of the sediment at different positions down the beach profile.

Bulk samples of sediment (surface and sub-surface) were taken on the river, delta and coast to achieve objective (c). Surface samples were taken from random 0.5m x 0.5m plots dug to the depth of the largest surface clast at each sample location (Figure 4.1). Sub-surface samples were extracted from the same plots.

Offshore sediment characteristics (objective (d)) were ascertained from surface seabed samples collected by Babbie Dobbie Ltd. (1994) and borehole logs from samples collected 100-200m offshore at Kingston (Moray Council 1999 unpublished data) (Section 3.5.2.2). In addition, a record of whether the surface seabed was gravel or sand was compiled during the offshore survey in August 1999.

4.5 Medium-term changes in fluvial and coastal planimetry

The 1:10 000 geomorphological maps of the lower River Spey (Figure 3.14) and Spey Bay (Figure 3.21) were used as bases to assess planimetric channel change and shoreline change over ca. 100 years. Channel change was determined from OS maps dated 1870, 1903 and 1971. In addition, historical vertical change was inferred using cross-sections surveyed in 1967 and 1889 (Lewin and Weir 1977) and comparing them to cross-sections surveyed herein at similar locations.

OS maps of Spey Bay dated 1870 and 1970 were compared to determine historical shoreline change for use in sediment budget calculations. The distance between the MHWS and MLWS at each beach profile location (every ca. 500m along the shore) was measured and used to create triangular beach profiles at each map date. The change in these over time was determined between 1870 and 1970 and erosional and accretional sections of the beach identified. Beach gradients at each point were calculated allowing beach steepening or flattening between the two dates to be determined. More recent changes in beach profile

were determined based on the profile network surveyed in this study (Section 4.3.1) at similar points along the beach to those profiles determined by the map analysis.

4.6 Survey spacing analysis

The spatial density of survey points on a river bar or beach profiles along a coastline influences the accuracy of volume and volume change calculations made using the data (Section 2.1.1 and 2.1.2). A high density of points or beach profiles will produce a more realistic estimation of the actual volumetric changes that have occurred. However, this necessarily requires a far greater field effort and may involve a trade-off of either limiting the spatial extent of the site or reducing the frequency of re-survey. A crucial part of this study was analysis to assess the effect of survey point and beach profile density on the accuracy of volume and volume change calculations, with the aim of selecting a survey density that produces estimates to within a level of accuracy suitable for this study. This analysis is central to assessing the accuracy of the sediment budget calculations.

4.6.1 River bars and subaerial delta

4.6.1.1 *Digital elevation model construction*

The terrain modelling package, Surfer for Windows (Golden Software 1994) was used to generate digital elevation models (DEMs) of the bar and delta surfaces. This package supports several interpolation options (summarised in Dixon et al. 1998), the choice of which depends on the type and use of the data. Kriging is used in this study as it has certain geostatistical optimal properties (see Davis 1986; Oliver and Webster 1990 and Cressie 1991 for further discussion) and, in particular, allows a value of the measurement error (or the micro-variance of the surface) to be input during grid interpolation. Kriging can thus be used as a smoothing interpolator to allow for the micro-variation of bar topography, as recommended by Lane (1998) and kriging has been shown to provide accurate estimates of volume differences from DEMs (e.g. Hicks and Hume 1997).

The so-called ‘nugget’ effect within kriging can be used to specify the errors of the data collection and is made up of the error variance (measurement errors) and the micro variance (small scale structure) (Cressie 1991). The measurement error can be estimated using the D_{50} of the bar surface sediment, as re-surveying on each survey point on a bar is likely to yield an elevation to within $\pm D_{50}$, depending on exactly where the survey pole is placed. For this analysis, the standard deviation (σ) of each survey measurement is estimated as $0.5D_{50}$ giving the error variance estimate of:

$$\text{Error Variance} = (0.5D_{50})^2 = 0.25D_{50}^2 \quad (4.1)$$

Field tests established the validity of this relationship.

Grid files were interpolated for each bar and delta surface using kriging and specifying the appropriate nugget effect using equation 4.1. An initial grid spacing of 2.5m was used. The calculation of bar and delta storage volumes together with volume changes between epochs is of importance in this study. Surfer uses three methods to calculate volumes or volume differences between grid surfaces (trapezoidal, Simpson's and Simpson's 3/8 rule; refer to Golden Software 1994 for details). The mean of the three results is taken as the best estimate of the true volume (cf. Hicks and Hume 1997).

The use of denser grids provides a more accurate volume calculation, particularly around the bar edges. The grids were expanded by cubic spline interpolation of two extra grid nodes between existing nodes in the original grid file. These, denser, smoother grids are used for all bar and delta volume calculations. Although this appears artificial, it preserves the existing grid nodes and is helpful for calculating differences between DEMs. Grid files created within Surfer are rectangular so blanking is used beyond the bar limits. The e, n coordinates defining the water's edge of each bar were used to define the boundary, the area outside of which is excluded from subsequent volume calculations.

4.6.1.2 Generation of bar surface changes

To determine the errors due to increased survey point spacing on volume change estimation, two new bar surfaces were artificially generated: one bar with the same topographic variation as the original bar, but of increased elevation ($z + 0.5$) and one bar with extreme topographic variability, but of the same dimensions as the original bar (z random).

4.6.1.3 Increasing the spacing of survey points

To determine the effect of increasing survey point spacing, the data for original and generated bars were progressively filtered (i.e. every 2nd, 3rd, 4th etc. data point were removed from the input data files prior to DEM interpolation). For example, two grid surfaces were generated using every 2nd data point only, each interpolated using only half of the original survey data. Storage volume and volume changes were calculated with these grid files and compared to those computed using all the data (referred to as the true volume or true volume change).

4.6.1.4 Base levels for volume storage calculations

To estimate sediment storage volumes reliably requires a realistic datum for volumetric calculations. River bar volumes were calculated using planar, sloping base levels, defined by a regression surface of the lowest level of the bed at 5 surveyed cross-sections in the reach with the following equation:

$$z(e, n) = 76.64 - 0.00238n - 0.00086e \quad (R^2=96.8\%, p<0.05) \quad (4.2)$$

The northing (which is roughly equivalent to the downstream distance) is the only significant variable in the regression. The easting is retained, however, as it allows a cross-stream component to be included.

Volume *change* at the delta is quantified using a horizontal base level at LWMS, which is below the lowest point of survey as this is suitable to compare *changes* in volume over time (Deruig and Louisse 1991). This avoids any confusion which may arise when combining the closure depth to the seaward and the depth of activity (i.e. lowest point of channel bed) to the landward of the delta storage zone. However, as the depth of the base level affects the relative difference between storage volume estimates as survey point spacing is increased, a horizontal base level which is equal to the lowest elevation of the original survey data was used for spacing analysis. This should minimise the effect of the base level when comparing the relative differences between volumes calculated with increased survey point spacing.

4.6.1.5 Results

(a) Sediment volumes

Sediment storage volumes were calculated for six bars and part of the Tugnet delta (Table 4.1) using base levels calculated as above. The effects of increasing survey point spacing on storage volumes are shown in Figure 4.2a-g. Increased spacing causes the calculated volume to deviate from the true volume and nearly always reduces volumes (Figure 4.2a-g). There is a wide disparity between volumes calculated for each particular spacing, depending on the actual survey data used to interpolate the grid. This highlights the sensitivity of the volumes (and the interpolated grids) to individual data points. Results from bar b and bar 5 are discussed in more detail.

The largest bar (bar b) has a total true storage volume of 94 565m³ (Table 4.1 and Figure 4.2f). This true volume is calculated from a DEM interpolated using an initial survey point spacing of ca. 4m, which may not itself be small enough to yield a realistic representation

of the surface topography. This spacing was chosen as an initial spacing of, for example, 1m is unrealistic, requiring a field survey of 52 630 points. The effect on estimated sediment storage volumes as spacing is progressively increased from 4 to 16m can be assessed. As spacing is increased calculated volumes begin to deviate from the true volume, and in the majority of cases the volume is under-estimated (Figure 4.2f). It should be noted that while the absolute volume error is often large (e.g. at a spacing of 16m the true volume is under-estimated by 2 114 m³) this translates into relatively small percentage errors (in this case 2.2%). This highlights the point that for large, macro-scale storage features larger absolute errors (and thus increased survey spacing) may be acceptable given that the error, as a percentage of the total storage volume, remains low.

bar no.	true storage volume (m ³)	Initial mean survey point spacing (m)
1	1597 ± 0.39	3.91
6	2459 ± 1.46	3.73
7	3210 ± 1.29	4.40
5	7683 ± 6.03	4.71
2	8490 ± 8.37	5.19
b	94565 ± 6.5	4.67
part of the delta	2710 ± 1.4	2.52

Table 4.1: Spey bar and delta true storage volumes. The true storage volume is calculated using all data with the initial spacing as shown. River bars are displayed with storage volumes in ascending order.

Bar 5 (Figure 4.3) has a true storage volume of 7683m³ calculated using a survey point spacing of ca. 4m. Again similar reservations exist as to how realistic a representation of the surface topography, and hence volume, this is. With an increase in spacing the volumes deviate from the true volume (Figure 4.2d) again generally being under-estimated (by nearly 8% at an 18m spacing). However, in some cases, the volume can be estimated to within 2.5 m³ (0.03%) of the true volume with a survey point spacing of 18m (using different points) (Figure 4.2d). While the storage volumes are almost identical, the DEM created with this 18m spacing has a very different surface topography (Figure 4.4) than that created using all data (Figure 4.3). How effective such a spacing will be at determining changes in sediment volumes is debatable and will be investigated later.

The average volume errors as spacing is increased (expressed as percentages of the true storage volumes) for all bars highlight the differences between large and small storage features (Figure 4.5). For example, increasing the survey point spacing on bar b, the largest storage feature, has little effect on the estimation of total sediment storage in percentage

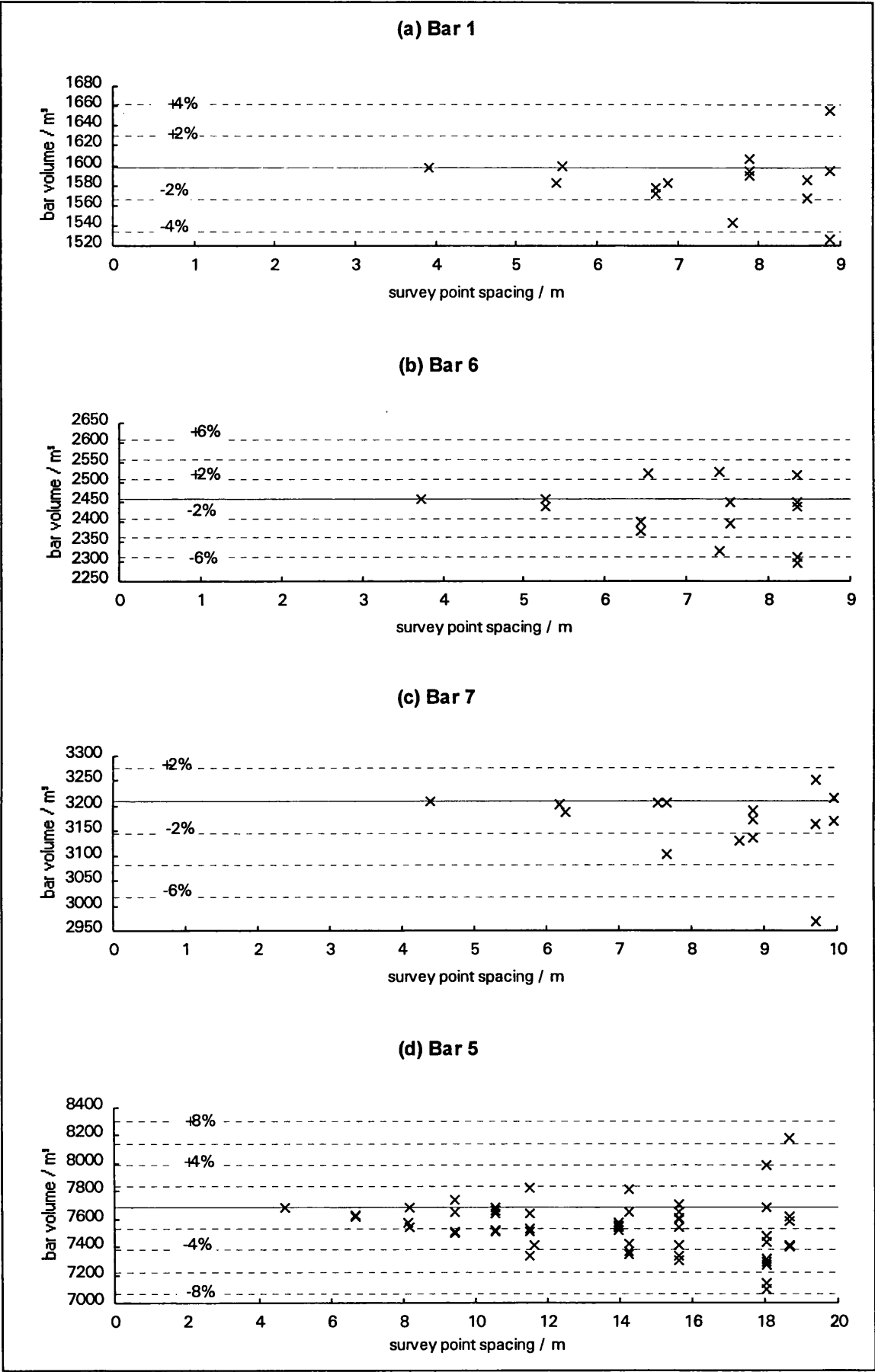


Figure 4.2: Volume calculations as spacing of survey points is increased (a) Bar 1 , true volume = 1597m³ (b) Bar 6, true volume = 2459m³ (c) Bar 7, true volume = 3210m³ (d) Bar 5, true volume = 7683m³.

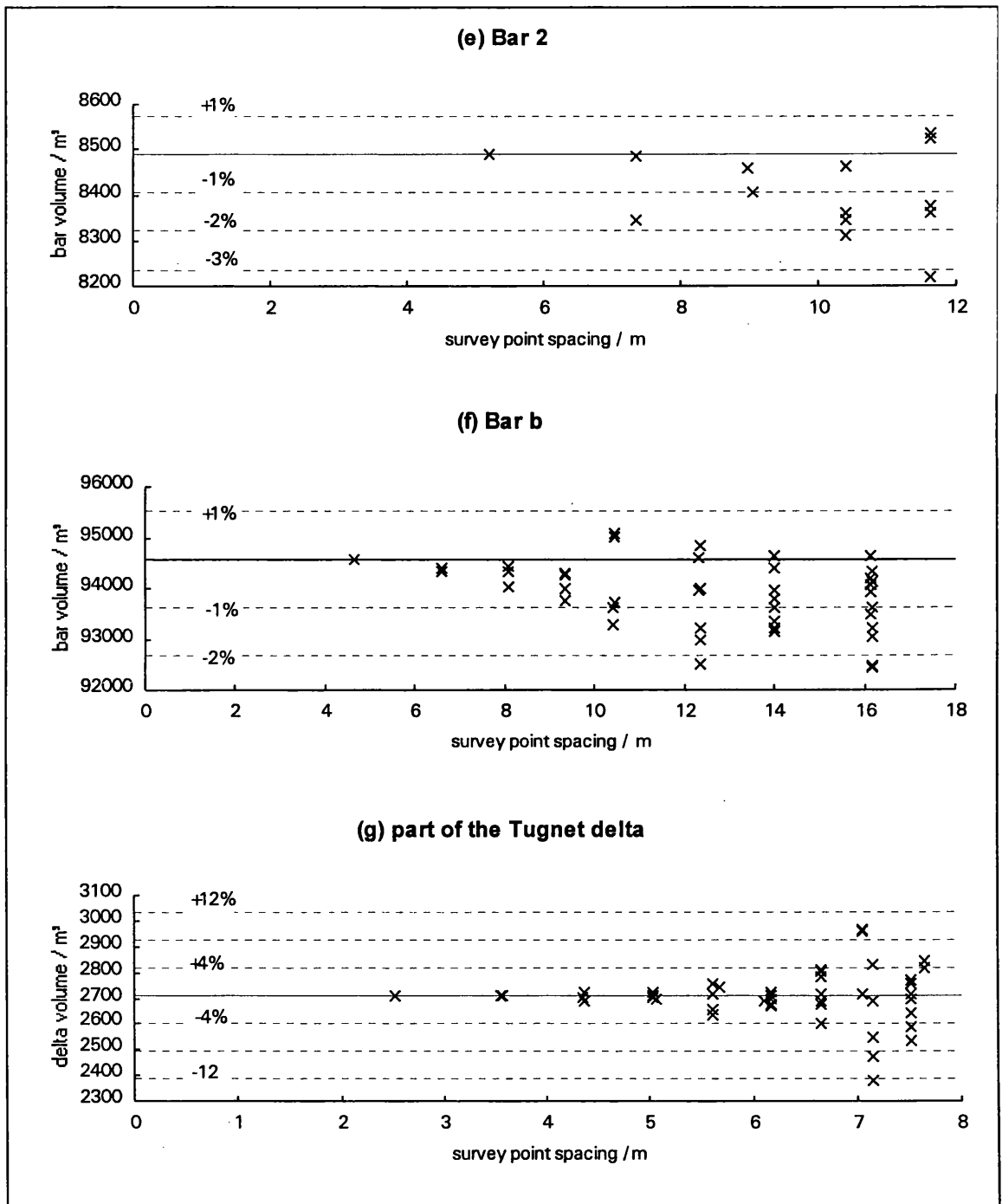


Figure 4.2: Volume calculations as spacing of survey points is increased, continued (e) Bar 2, true volume = 8490 m³ (f) Bar b, true volume = 94560 m³ (g) Delta, true volume = 2710 m³.

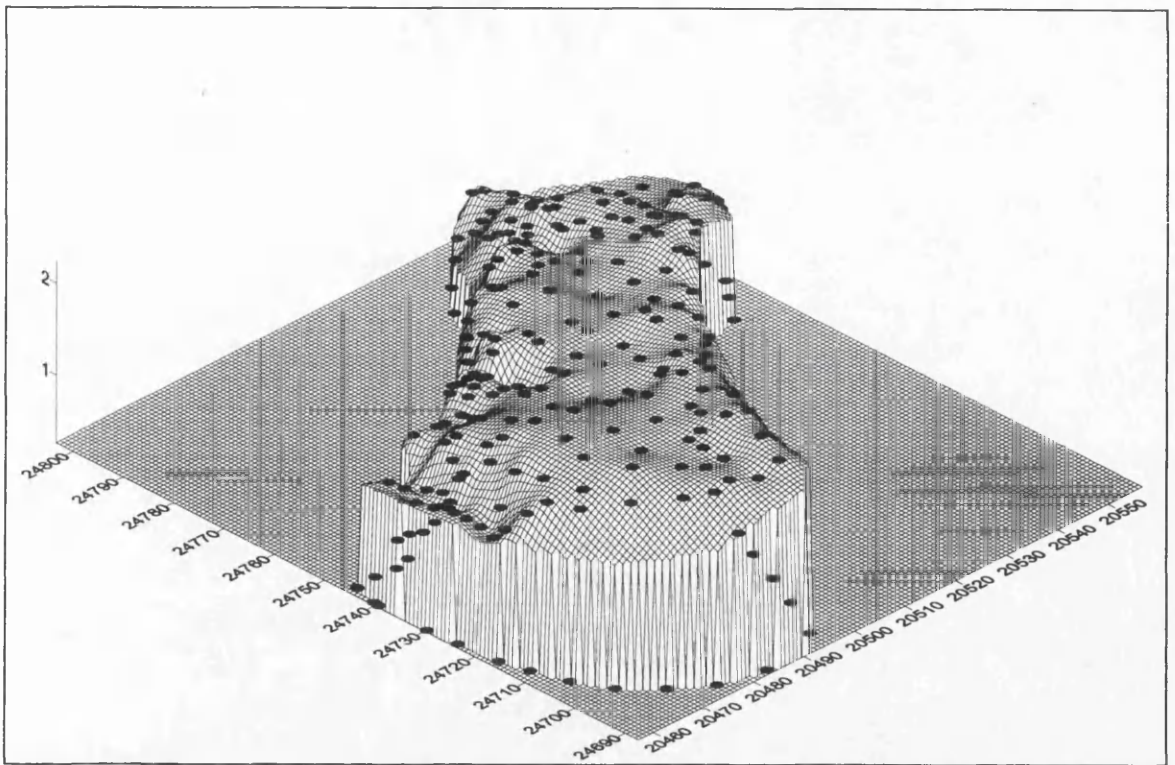


Figure 4.3: Bar 5 interpolated using all data has a true storage volume of $7683 \pm 6 \text{ m}^3$.

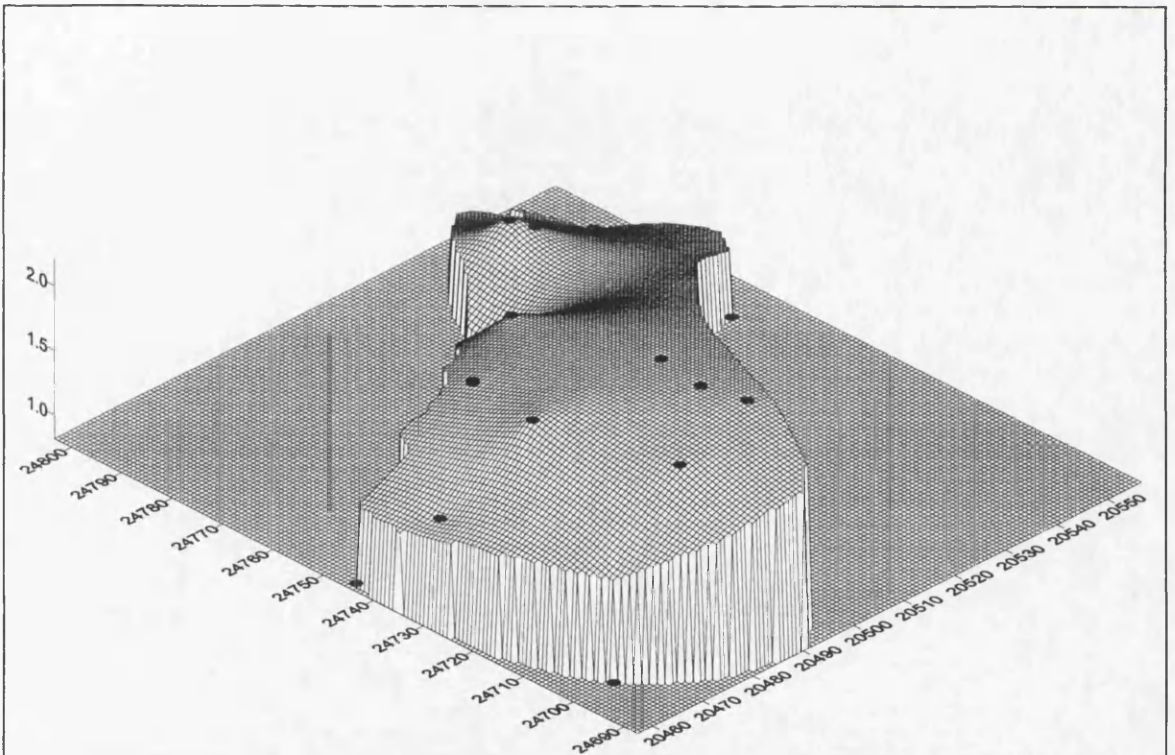


Figure 4.4: Bar 5 interpolated using every 15th survey point remaining (i.e. a survey point spacing of 18m). The storage volume calculated from this DEM is $7680 \pm 6.5 \text{ m}^3$ (only 2 m^3 different from the true volume). Note, this is only one of the several realisations at this spacing.

terms (the average volumetric errors are never more than 1% of the true storage volume with spacing up to 16m). In contrast, the smallest storage feature (bar 1) is more sensitive to increasing survey point spacing (Figure 4.5). This highlights the need to determine the survey spacing on the basis of the scale of the feature and the acceptable errors required to achieve a particular study aim.

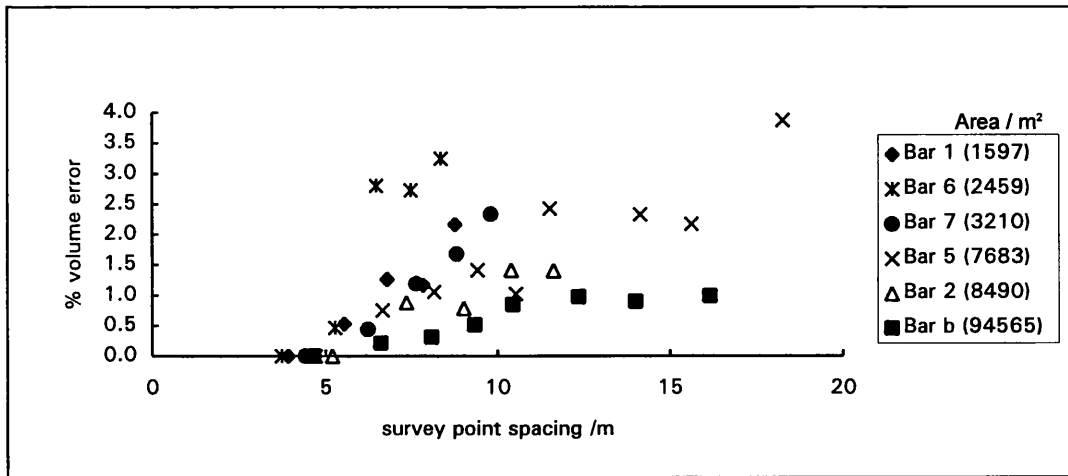


Figure 4.5: Mean absolute volume errors (%) as survey point spacing in all bars is increased. Note the sediment storage volume percentage error is less for large storage features (bar b).

(b) Sediment volume changes

To investigate the effect of survey density on the estimation of bar volume change two new bar surfaces were generated for bar b and bar 5 (see Section 4.6.1.2). Volume changes for each bar were calculated using the generated bar surfaces as the upper grid surfaces with the original bar surface as the base. Volume changes were calculated firstly using all survey data (in both the upper and lower surface) and then progressively increasing the spacing.

For the larger bar b, the total volume change with a uniform elevation increase of 0.5m is 26 514 m³ (Figure 4.6a). As survey point spacing is increased the error associated with the volumetric change calculation also increases (Figure 4.6a). At spacings of less than 10m the volume change lies within $\pm 2.5\%$ of the true volume change, representing volume errors of up to 660 m³. While this may be considered a large volume of sediment to ‘miss’, it may not be significant given the magnitude of the overall volume change. As spacing increases beyond 10m the mean percentage error increases substantially (Figure 4.6a); at a spacing of 12m the volume error can be under-estimated by up to 8.7% (2316 m³).

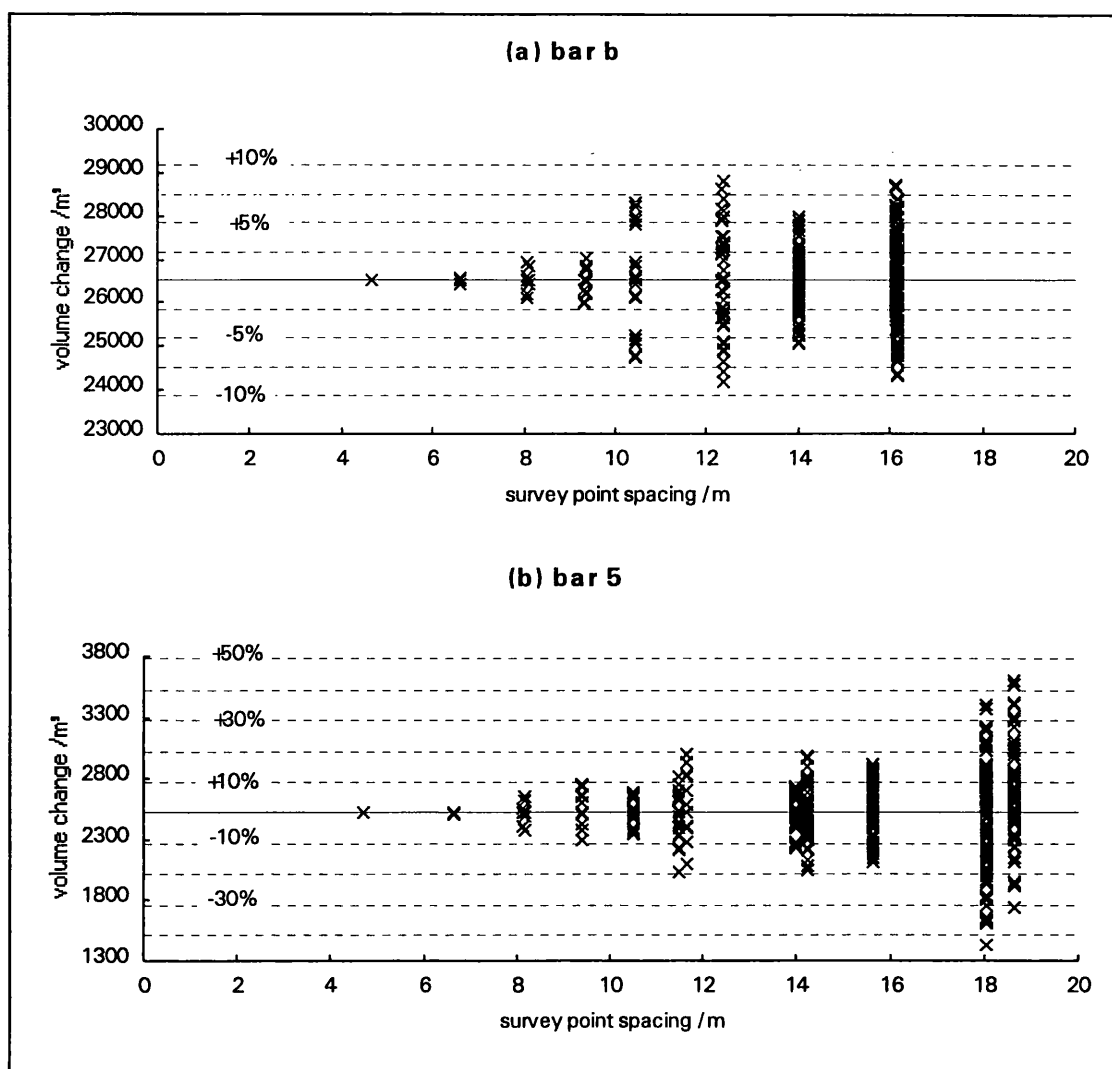


Figure 4.6: Volume change when bar elevation is uniformly increased by 0.5m for (a) bar b (b) bar 5.

Using all data for bar 5 the true volumetric change is estimated as $2\,519\text{ m}^3$ (Figure 4.6b). This change is much smaller than that for bar b, and so the percentage errors in volume calculations as spacing is increased are substantially larger. With spacings of less than 10m, the change can be detected to within $\pm 10\%$ (252 m^3), although as spacing increases to over 11m the errors double (Figure 4.6b) and at a survey point spacing of 18m the volumetric change can be under or over-estimated by up to 43% (1083 m^3).

The mean absolute percentage errors as spacing is increased for both bars show that the percentage error is a function of the magnitude of actual volume change (Figure 4.7). The percentage errors are smaller when detecting volumetric changes of greater magnitude (e.g. an error of 900 m^3 represents only a 3.3% error for bar b but is a 36% error for bar 5). Thus, to detect small volumetric (and thus topographic) changes to within a given error, closely spaced surveys may be required, while to detect much larger changes a less dense survey will often suffice.

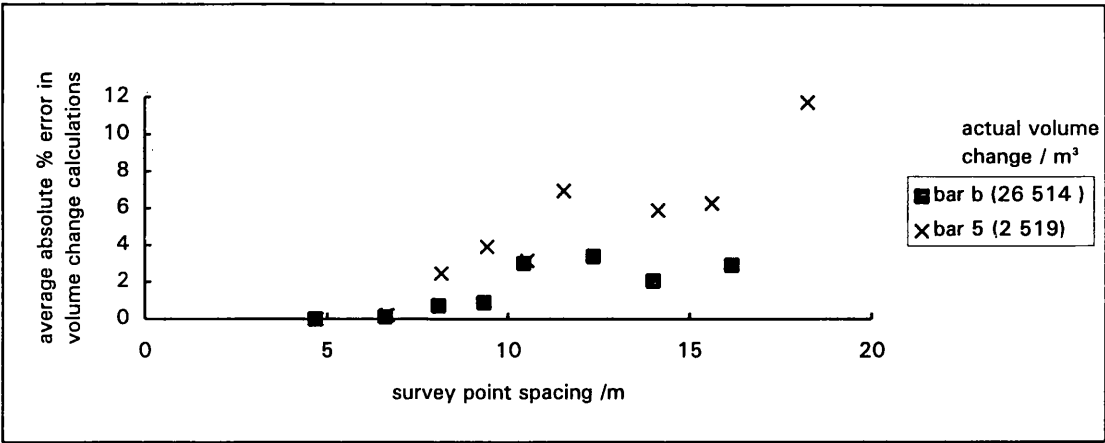


Figure 4.7: Average absolute volume change errors (expressed as a % of the actual volume change) as spacing is increased for bars experiencing an elevation increase of 0.5m

For a randomly generated bar topography, errors in the estimation of volume change are considerable as survey point spacing increases (Figure 4.8). Large percentage errors are associated with the detection of small volumetric changes (e.g. bar 5).

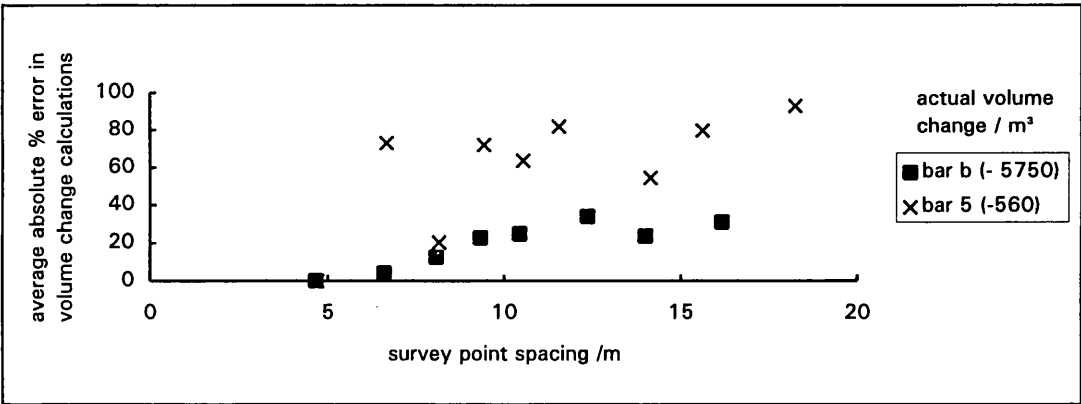


Figure 4.8: Average absolute volume errors (expressed as a % of the actual volume change) as spacing is increased for bars with a randomly generated topography.

4.6.1.6 Conclusions and implications for survey point density

The above analysis shows how errors in volume change calculations are a function of survey point density and highlights the need to chose the most suitable density to fulfil a particular study aim. For example, a much denser survey is required to accurately quantify volumetric changes compared to that required to quantify the absolute volumes of sediment in storage.

The survey point density required to quantify storage volumes to within a given error range varies with the size of the storage feature (Figure 4.9a). Figure 4.9a shows the error as survey point spacing is increased where:

$$\% \text{ error} = ABS \left(\left[\frac{\text{volume calculated with increased spacing}}{\text{true volume (using all data)}} \times 100 \right] - 100 \right) \tag{4.3}$$

The smallest storage features (e.g. bars 1 and 6 and the delta) are the most sensitive to increasing survey point spacing (Figure 4.9a) with a spacing of ca. 6m causing deviations from the true volumes of $\pm 2.5\%$. In contrast, spacings of up to 16m can quantify the volume of sediment in storage in bar b (the largest storage feature) to within $\pm 2.5\%$ of the true volume. Overall, a survey point spacing of 6m is recommended to estimate the bar volume to within $\pm 2.5\%$ of the true bar volume in all cases. The delta is highly sensitive to any increase in spacing over 6m (Figure 4.9a). If spacing is made independent of bar area the bar volume can be calculated to within $\pm 2.5\%$ of the true volume when the spacing/ $\sqrt{\text{area}}$ ratio is less than 0.15 (Figure 4.9b). Percentage errors increase substantially when this ratio is exceeded.

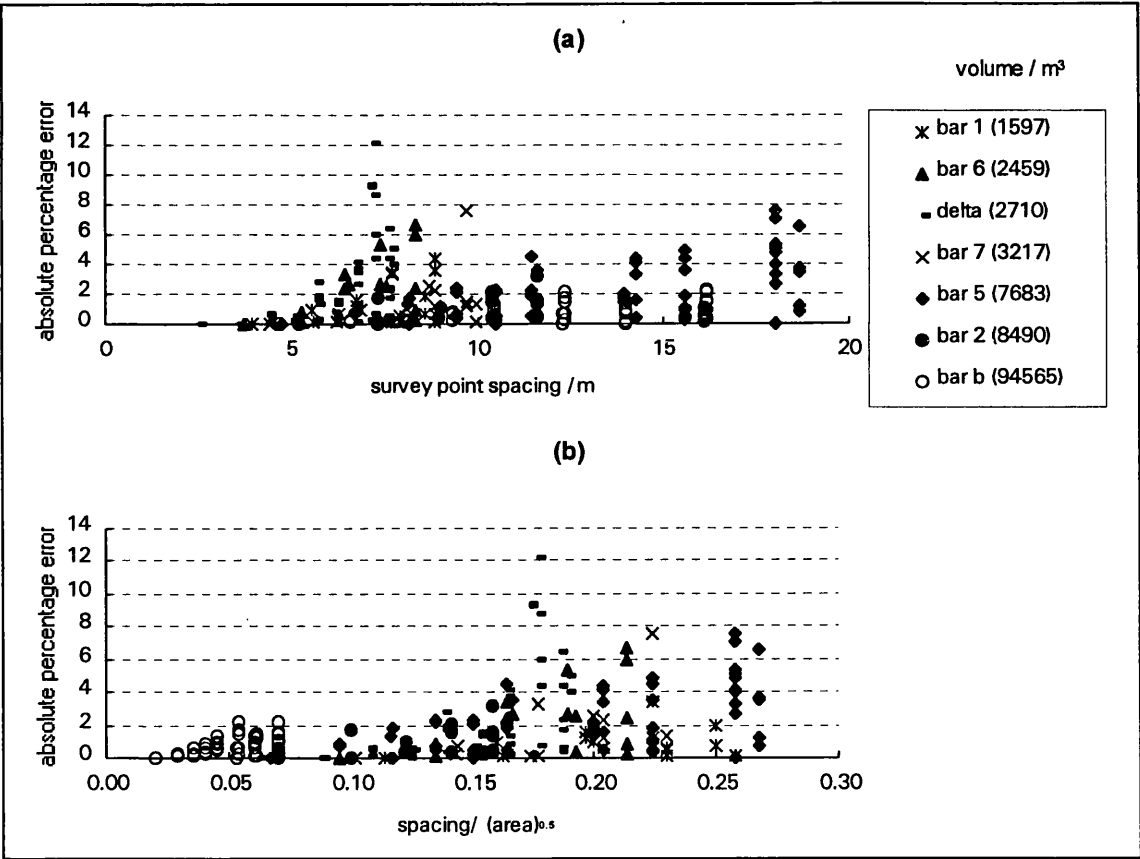


Figure 4.9: Absolute percentage errors in storage volume calculations as spacing is increased in all bars and the delta (a) with survey point spacing on the x-axis (b) with $\frac{\text{spacing}}{\sqrt{\text{bar area}}}$ on the x-axis.

To calculate volume changes the density of points must be appropriate to resolve sediment volume changes to within the magnitude of those actually occurring. To detect very small volumetric changes to within a given error, dense surveys are required. For larger

volumetric changes a less dense survey will often suffice. For example, with a 10m survey point spacing large volumetric changes in bar b (26 500 m³) are calculated to within $\pm 2.5\%$ of the true volume change (Figure 4.7) but for smaller changes (-5 750 m³) errors increase to $\pm 25\%$ (Figure 4.8).

4.6.2 Beach profiles

To accurately quantify sediment volume changes along a shoreline using beach profiles the spacing of these profiles is critical (Section 2.1.2). Profile locations should be chosen such that each profile is representative of the change in the sediment cell it represents (i.e. representative of the change in shoreline from it to a half distance to each adjacent profile). While several studies (e.g. Grove et al. 1987; Foster et al. 1994; Hicks et al. 1999) have used beach profiles to quantify volumetric gains and losses, none have undertaken error analysis to identify the appropriate spacing of profiles for accurate cell volume computations. Most studies have defined profile spacing with little methodological justification. As this study requires accurate volumetric changes, analysis was undertaken at two scales (100m and 10m) to determine the appropriate spacing of profiles and the errors associated with given spacings.

4.6.2.1 Data collection 1 (100m profile spacing)

Eight shore-normal profiles were surveyed with a longshore spacing of 100m (Figure 4.10). These profiles, between 1.6 and 2.3km west of the mouth of the Spey, were chosen as representative of beach profile variation within Spey Bay.

4.6.2.2 Storage volume calculations

The area under each profile was calculated down to the maximum depth surveyed (-0.899m OD) and the reference distance was taken as the distance from the main beach crest (Figure 4.10). The area calculated is the triangular area beneath the profile and represents the total volume of sediment in storage. The volume at a given profile is given in m³/m of shoreline (i.e. it is directly equivalent to an area). Volumes can be calculated between profiles assuming that the area at a profile is representative of the distance between it and the half distance to each adjacent profile, using:

$$Volume = \frac{A_i + A_{(i+1)}}{2} L_{(i, i+1)} \quad (\text{equation 2.11, chapter 2})$$

where A_i is the area at profile i , $A_{(i+1)}$ is the area at an adjacent profile and $L_{(i, i+1)}$ is the distance between the two profiles. This assumes that the profiles are parallel and of equal length. It also assumes gradual monotonic change in elevation between profiles (i.e. no

hills or troughs between the profiles). Violation of these assumptions can introduce significant error and systematic bias.

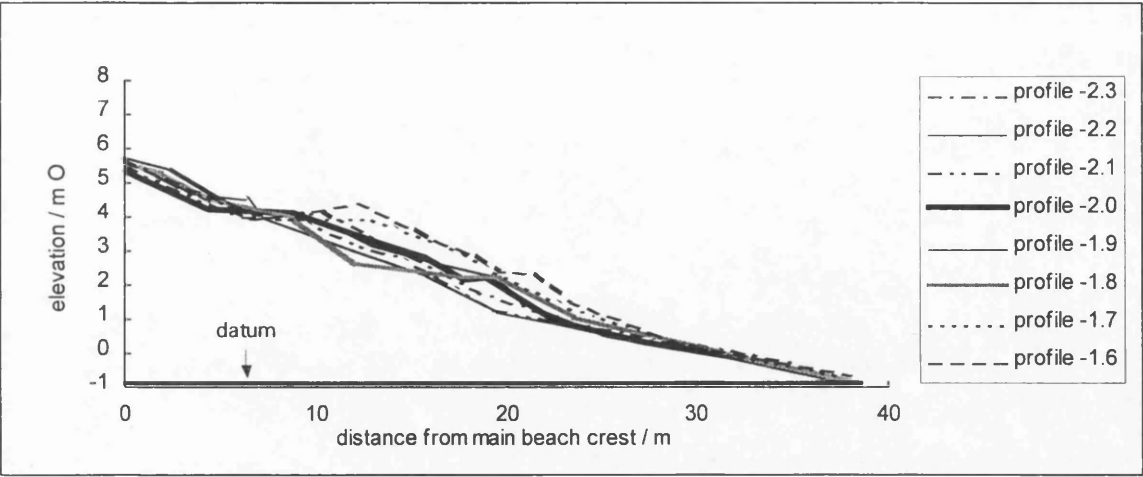


Figure 4.10: Closely spaced beach profiles. Profile -2.0 (shown in bold) would be taken as the representative profile if a spacing of 500m had been used.

4.6.2.3 Profile spacing analysis 1

Volumes were calculated for each 100m cell of the surveyed shoreline using equation 2.11 (Table 4.2). These volumes were assumed to represent the true beach volumes.

Profile	Area (m ³ /m)	Spacing (m)	Volume (m ³) (rounded to 3 s.f.)
-2.3	114.9		
		104.1	11 600
-2.2	107.5		
		100.8	10 900
-2.1	108.4		
		102.7	11 300
-2	111.7		
		99.0	11 200
-1.9	114.8		
		100.8	11 500
-1.8	112.7		
		97.8	11 500
-1.7	122.8		
		98.3	12 300
-1.6	128.1		

Table 4.2: Beach profile areas (m³/m) and cell volumes (m³) for each ca. 100m wide cell.

Profiles were successively removed and the volume calculations repeated, so increasing the inter-profile spacing to ca. 200, 300, 400, 500, 600 and 700m respectively. These volumes were then compared to the best estimate of the true beach volumes for that cell (calculated by addition of all the 100m cell volumes contained within the larger cell). The differences

between these volume estimates give estimates of error sensitivity to beach profile spacing, where error is:

$$\% \text{ error} = \left[\frac{\text{volume calculated with increased spacing}}{\text{true volume (with 100m spacing)}} \times 100 \right] - 100 \quad (4.4)$$

As profile spacing is progressively increased the calculated volumes deviate from the true volumes and tend to over-estimate them (Figure 4.11). Increasing the spacing from 100m to 200m produces deviations from the true beach volume of less than 3%. This percentage loss does not increase substantially up to 500m spacing. However, at spacings greater than 500m the beach volumes can be over-estimated by up to 6.5% (Figure 4.11). In order to maintain beach storage volume calculation errors to within $\pm 2.5\%$ (as with the river bars, Section 4.6.1.6) a 500m profile spacing is used in this study. Less densely spaced profiles (e.g. 700m) cause unacceptable deviations from the true volumes (up to 6.5%).

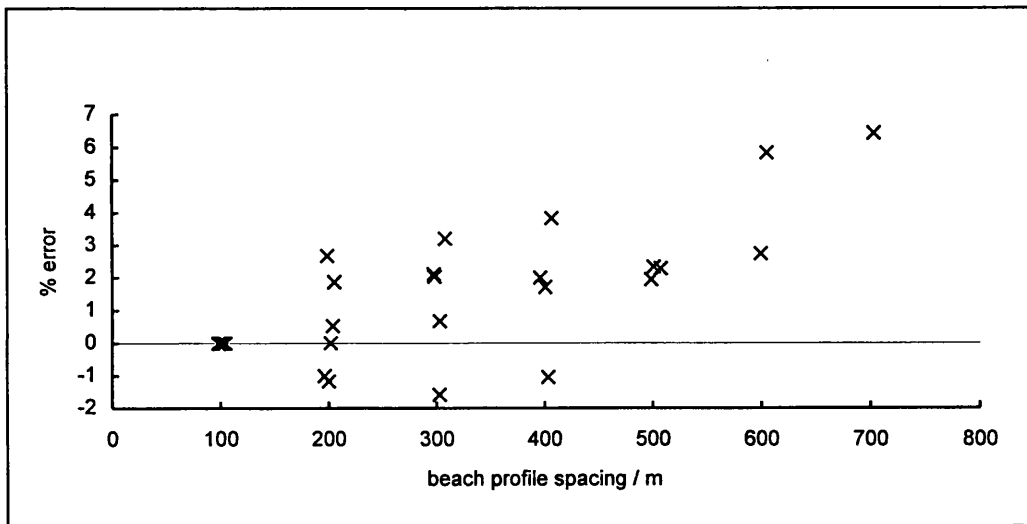


Figure 4.11: Volume errors as a function of profile spacing. Note the volume calculated using a profile spacing of 100m is assumed to represent the true beach volume.

Errors, when expressed as percentages of the true volume, are relative to the magnitude of the true volume or true volume change. In the above analysis the true volumes were calculated by calculating the triangular area under each profile, using a horizontal base level, and thus are relatively large compared to the calculation errors (giving small percentage errors). To assess the effect of base levels on error, a sloping datum parallel to the mean beach gradient but lying below all survey data was also used to calculate the profile area. The true beach volumes using this method are smaller and ultimately depend on the depth of the sloping base level. As profile spacing is increased the absolute error (in m^3) remains approximately the same no matter what base level is chosen (Figure 4.12), however the choice of base level significantly affects the percentage errors.

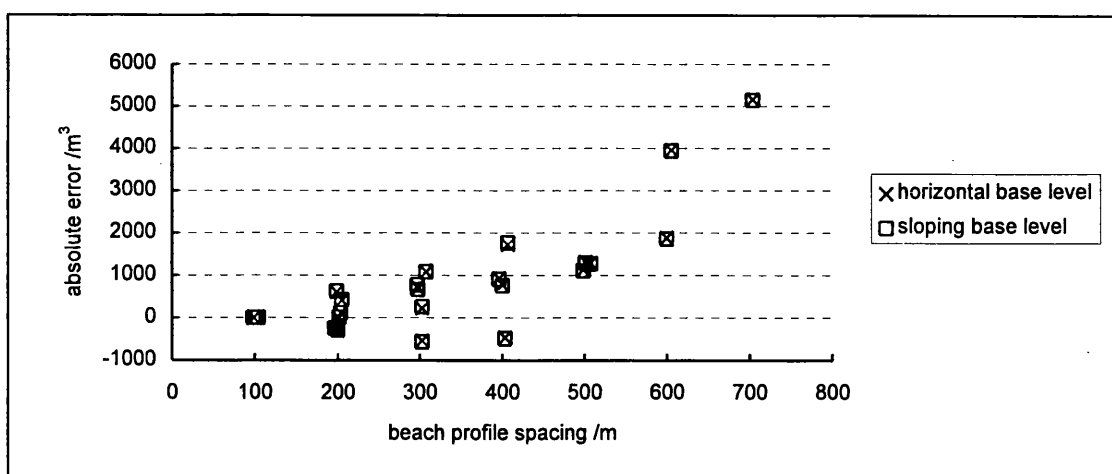


Figure 4.12: Absolute volume errors versus profile spacing. The base level used to calculate the area at each profile has little effect on the absolute errors (m^3) but significantly affects the percentage errors. The same absolute error at 700m represents a 6.5% error using the horizontal base level (true volume is $80\,280\text{ m}^3$) and 19.1% error using a sloping base level (true volume is $26\,630\text{ m}^3$).

4.6.2.4 Dense survey of a section of active beach

The above analysis assumes that the storage volume calculated using a beach profile spacing of 100m represents the true volume within each 100m cell and accurately quantifies this volume with no error. Thus the $\pm 2.5\%$ error in volumetric calculations at a 500m spacing only holds if there is no error involved when using a 100m profile spacing. To investigate this potential oversimplification, analysis of a short, densely surveyed, section of the beach was undertaken.

4.6.2.5 Data collection 2 (10m profile spacing)

A 200m section of the active beach was surveyed with an average survey point spacing of ca. 7m (Figure 4.13). The DEM was constructed as described in Section 4.6.1.1. Closely spaced beach profiles at a longshore spacing of 10m were extracted from this DEM. Shore-normal profile lines were defined by two co-ordinates in a blanking file and the slice command within Surfer was used to generate cross-section data along these profile lines. As blanked data was not included in the analysis, the beach profiles begin at a common reference point (i.e. the main beach ridge crest) and extend to the lower foreshore to the maximum depth surveyed. The profile data is output by Surfer in the form of horizontal distance and elevation at every point where the profile line intersects a grid line. Areas under each profile (m^3/m) were calculated from this data. Volumes of sediment contained in each 10m cell were then calculated using equation 2.11.

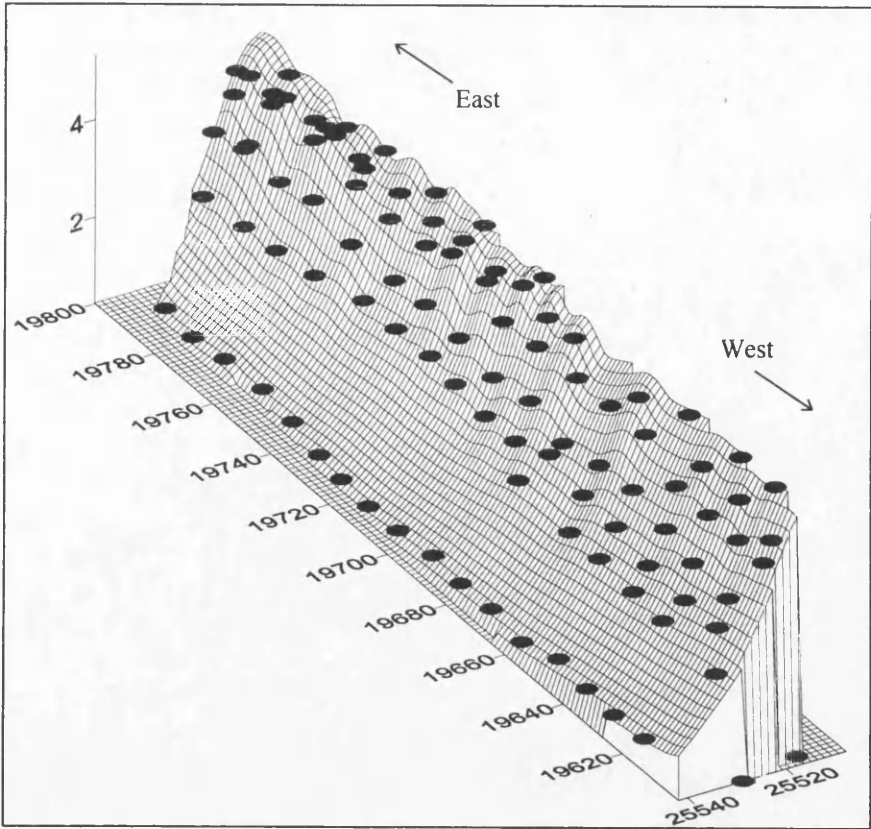


Figure 4.13: Survey of the active beach at Kingston. The active beach is defined by 106 survey points with an average spacing of 7.14m. Shore-normal profiles were extracted from this grid surface every 10m.

4.6.2.6 Profile spacing analysis 2

Volumes calculated for each 10m cell are considered to represent the true beach volumes. As before, profiles were successively removed from the volume calculations, increasing the spacing in 10m intervals to a maximum spacing of 180m. Errors as beach profile spacing is increased were calculated using equation 4.4.

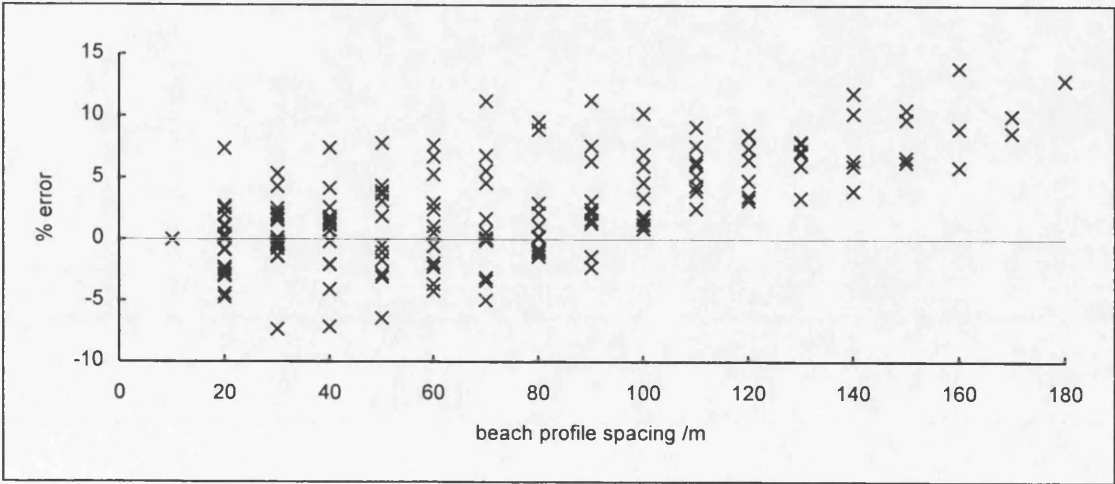


Figure 4.14: Volume errors as a function of profile spacing. Note the volume calculated using a profile spacing of 10m is taken to represent the true beach volume.

As spacing increases the calculated volumes begin to deviate from the true volumes and there is a tendency for volumes to be overestimated. Increasing the spacing to 20m causes the calculated volume to deviate from the true volume by up to 7% (+74 m³) (Figure 4.14). As profile spacing is increased the error progressively increases (Figure 4.14). With profile spacing greater than 100m the volume calculated is consistently overestimated, reaching almost 14% at a spacing of 160m. A profile spacing of 100m can overestimate the volume by as much as 566 m³ (10%) and the mean error with a 100m spacing is 222 m³ (4.1%).

4.6.2.7 Discussion

Results from both the closely spaced surveyed beach profiles (Section 4.6.2.3) and the profiles interpolated from the Kingston DEM (Section 4.6.2.6) were combined to determine the mean absolute errors in volumetric calculations associated with a particular spacing (Figure 4.15). Absolute errors are discussed here, as percentage errors are highly dependent on the choice of base level (Section 4.6.2.3).

The 10m spacing analysis suggests that storage volumes calculated using a 100m profile spacing have mean absolute errors of 222 m³ (4.1%). As the surveyed beach profiles have an initial spacing of 100m, the volumes calculated at this spacing (the true volumes of Section 4.6.2.3) have an inherent error of the order of ± 200 m³. Errors for spacings greater than 100m are calculated as:

$$e = \sqrt{(e_1)^2 + (e_2)^2} \quad (4.5)$$

where $e_1 = 222$ m³ and e_2 is the mean absolute error for a given spacing greater than 100m.

Errors appear to be acceptable up to a 500m spacing, but any further increase in profile spacing increases the errors substantially (Figure 4.15). A beach profile spacing of 500m has an associated error of ± 1260 m³ ($\pm 2.3\%$ of the true volume) which is considered acceptable for this study. A denser survey coverage (e.g. 100m) decreases the error to ± 222 m³ for each 100m cell (or ± 1110 m³ for a 500m cell). While this accuracy is ultimately more desirable for the computation of a sediment budget, the increased field time is not justified for such a limited increase in accuracy. To decrease field time by increasing profile spacing to say 700m is not justified as errors increase substantially to ± 5200 m³ ($\pm 6.5\%$ of the true cell volume). A 500m profile spacing is used and all volumetric calculations are quoted with the calculated uncertainty. Note, the actual volumetric changes (in m³) are likely to be smaller than in this analysis as it is the *change* in area at each profile that will be input into equation 2.11.

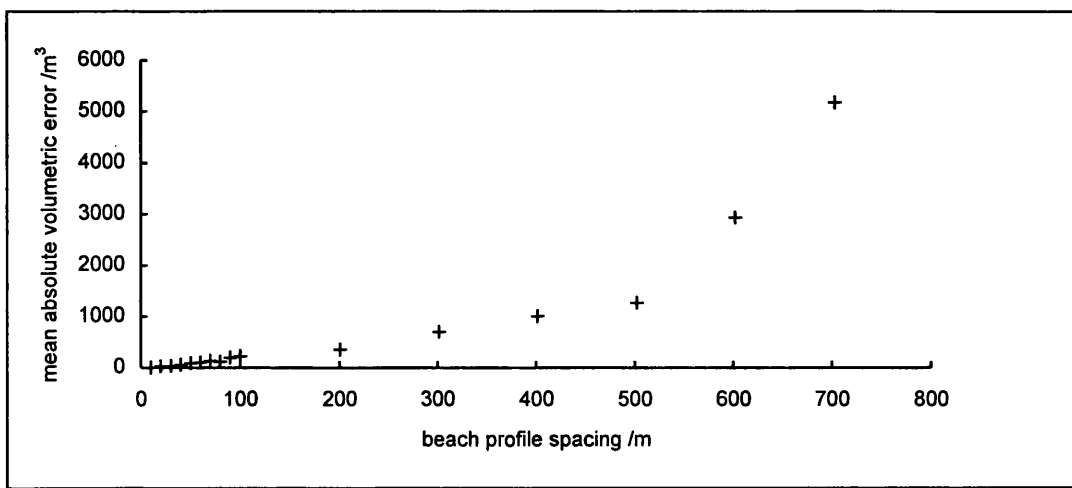


Figure 4.15: Mean absolute volumetric error versus profile spacing. A 500m spacing will quantify the volume of sediment in storage in that cell to within $\pm 1260 \text{ m}^3$.

4.6.2.8 Volume change calculations

As beach volume change will generally be significantly less than the storage volume, the above analysis was repeated using two surfaces of change (Figure 4.16): (a) an artificial profile generated by a regression through all profile -1.6 to -2.3 data (adjusted $R^2 = 0.965$) and; (b) data from a repeat survey of profile -2.0 (7/9/97).

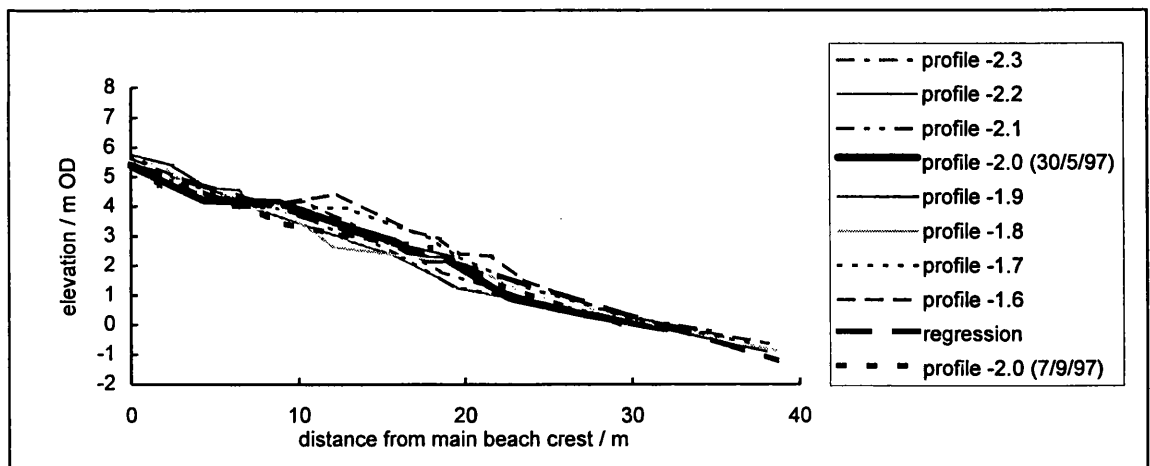


Figure 4.16: Beach profile change analysis. Two profiles were used to assess the sensitivity of beach volume changes to increasing profile spacing: a linear regression of all profile data and the actual change recorded when profile -2 was re-surveyed.

The area change at each profile was calculated and converted to volume changes for each beach cell (Table 4.3). Profile spacing was increased and volume changes calculated for successively wider cells. Using either surface of change, most profiles experience a combination of both cut (negative area change) and fill (positive area change) at various points down the profile (Figure 4.16). The net area change at any given profile is generally small and may be either positive or negative (Table 4.3) leading to large absolute errors as spacing increases (Figure 4.17).

Profile	Area Change (a) (m ³ /m)	Area Change (b) (m ³ /m)	Spacing (m)	Volume Change a. (m ³)	Volume Change b. (m ³)
-2.3	-2.0	11.7			
			104.1	-85	494
-2.2	0.4	-2.2			
			100.8	-407	-177
-2.1	-8.5	-1.3			
			102.7	-701	34
-2	-5.2	2.0			
			99.0	-165	363
-1.9	1.9	5.4			
			100.8	-75	420
-1.8	-3.4	3.0			
			97.8	-145	784
-1.7	0.4	13.1			
			98.3	37	1544
-1.6	0.4	18.3			

Table 4.3: Beach profile area change and volume change calculations using two profiles of change a. regression and b. repeat survey of profile -2.0.

In percentage terms the errors can be high (Figure 4.18). In this case, a 500m spacing is associated with a mean percentage error of $\pm 118\%$ when estimating volume change. Again, it is re-stated that the percentage error is dependent not only on the absolute error (in m³) but also on the magnitude of the volume change, which in this example is very small. For large volume changes associated with major profile shifts (e.g. beach retreat caused by a major storm) the percentage error will be much smaller.

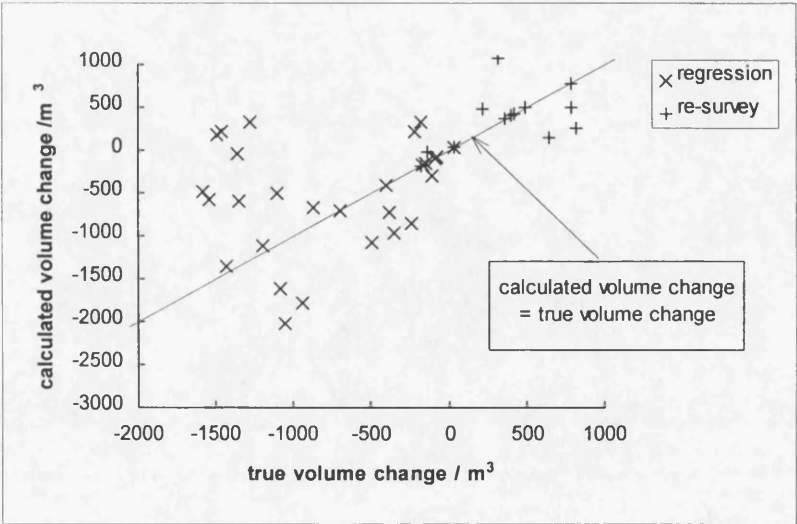


Figure 4.17: Volume change calculated as profile spacing is increased versus the true volume change. Both the regression and re-survey data are highly sensitive to increasing profile spacing as the calculated volumes deviate substantially from the true volumes.

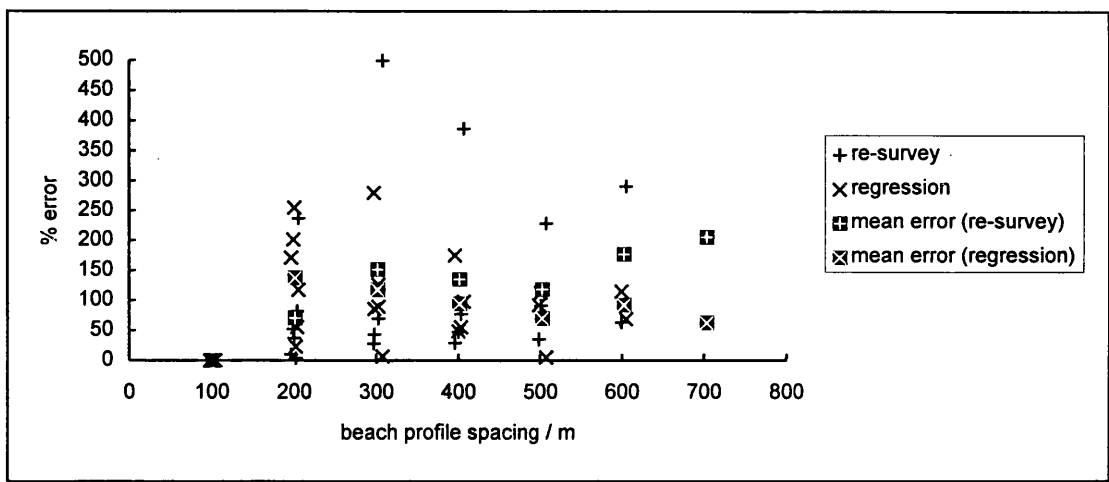


Figure 4.18: Volume change error versus profile spacing. The mean percentage error for each spacing is shown. Note there is no progressive increase in error as spacing increases and the mean % error actually decreases in some cases.

The quantification of beach volume change along a shoreline is highly sensitive to profile spacing. Errors of up to five times the true change can be introduced by merely trebling the spacing (Figure 4.18). These high errors are likely to be related to the scale of the change (Table 4.3) with very small changes of varying magnitude inducing large relative error. As the scale of the change increases the relative error will decrease. It follows that using beach profiles to detect small changes is problematic as the error will often be of greater magnitude than the true change.

To quantify volume changes along a shoreline using repeat beach profile surveys requires a sound geomorphological knowledge of the beach and the changes that are occurring. For example, if one profile experiences significant cut while the adjacent profile experiences significant fill it is important that the researcher understands what is going on in between. Is there a gradual monotonic change from erosion to accretion? If so, the technique is valid. However, if the entire cell is erosional except for a short section where the fill profile is located, any change quantified using this technique is invalid.

To minimise errors a profile spacing of ca. 500m was chosen (Section 4.6.2.7) and profile location was based on a thorough understanding of the geomorphology and the longer term trends of coastal adjustment (Section 6.1). During repeat surveys, the changes that occurred between profiles were observed and, if need be, extra profiles were surveyed at sensitive locations (e.g. the gravel-sand beach transition). Observations of geomorphological changes between profiles were taken into account when calculating cell volume changes, to ensure that the change is representative of what actually occurred.

4.7 Summary

In order to accurately quantify sediment storage and transfers in the fluvial, deltaic and coastal environments, a range of field methods have been selected that balance logistics with accuracy. A full error analysis of the sediment budgets and storage changes presented in Chapters 5, 6 and 7 is undertaken.

5. FLUVIAL AND DELTAIC SEDIMENT STORAGE AND TRANSFERS

In order to calculate the Spey Bay sediment budget, the storage and transfers of sediment in each of the fluvial, deltaic and coastal sub-systems are quantified individually using the methods outlined in Chapter 4. This chapter presents the fluvial and deltaic sediment budgets, with the coastal system presented in Chapter 6. The combined sediment budget for the whole of the Spey Bay system is synthesised and discussed in Chapter 7.

5.1 Medium-term fluvial change

Estimation of the rate of fluvial morphological adjustment is highly dependent upon the time interval over which the changes are observed (Inglis et al. 1988). Recent changes may be short-term adjustments within a longer term trend or may represent noise around a constant condition. To put the short-term changes investigated in this study (Section 5.2) into perspective, the medium-term record of channel change in the lower River Spey was analysed.

Lewin and Weir’s (1977) analysis of a series of maps show a general decrease in braiding in the lower River Spey since 1760 (Table 5.1). The nineteenth century maps indicate a different style of braiding than is present today with a greater number of larger bars, often without a single dominant channel (Lewin and Weir 1977). The braiding index, calculated based on the lengths of mid-channel islands and bars (Brice 1960), decreased from 7.4 in 1760 to 2.1 in 1995. However, there are problems associated with the use of maps since the procedures of surveying detail in the more recent metric maps differ from earlier methods: McEwen (1989) found that braiding was reduced in the Dee when plotted from metric maps and so the results must be viewed with caution.

Date	Map scale	Braiding Index
1995	1:10 000	2.06
1967	1:10 000	2.26
1889	1:2 500	6.50
1887	1:2 500	6.43
1882	1:2 500	5.32
1876	1:11 000	5.00
1760	1:21 000	7.40

Table 5.1: Braiding Index for the lower River Spey (adapted from Lewin and Weir 1977)

Consistent with the observed decrease in braiding, the active channel width has also decreased (Figure 5.1). The active channel width (which includes unvegetated bars and islands as well as the wetted channel) has decreased by 60%, from an average of 266m in 1870 to 108m in 1971 (Gemmell et al. 2000). Wider areas of active channel generally indicate areas of sediment storage (Church 1983), the locations of which have varied through time in the lower Spey (Figure 5.1).

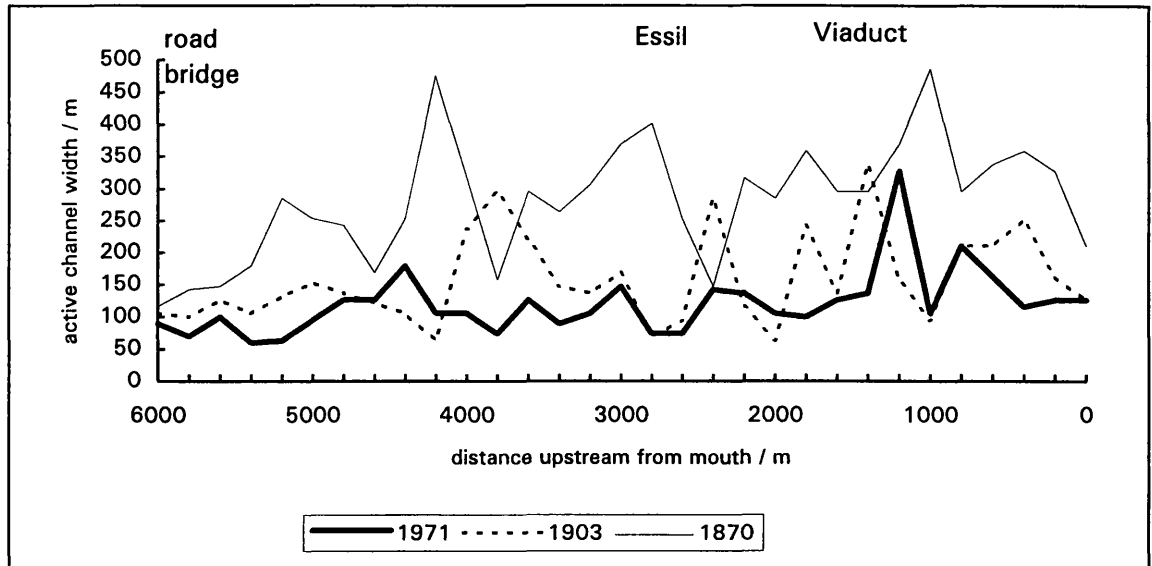


Figure 5.1: Variation in the active channel width of the lower River Spey (1870, 1903 and 1971)

In active river reaches, major adjustment occurs in response to high magnitude events (e.g. Anderson and Calver 1980; Ferguson and Werritty 1983; McEwen 1989; Warburton et al. 1993) as in many cases a major threshold has to be exceeded before areas of floodplain can be excavated or reworked. The Great Moray Flood of 1829 (Section 3.3.1) had a considerable geomorphic impact within the Spey basin, as large volumes of sediment were eroded and transported downstream (Lauder 1873). This impact was long-lived and has influenced the location of subsequent erosion, providing the initial access to sediment (Inglis et al. 1988). A network of flood channels which may date from the 1829 flood persists today (Figure 3.14) and are reoccupied during contemporary river spates (Section 3.3.2.1).

The active channel has occupied its present general course (within the scrub and woodland vegetated floodplain) since at least 1760, but within this area, the location and the number of channels have varied considerably (Lewin and Weir 1977). Large changes in channel planform have been observed in short periods, particularly near the river mouth, north of the Speymouth viaduct (Riddell and Fuller 1995). Channel changes mapped from Ordnance Survey maps dating back to 1870 (Figure 5.2, in sleeve) are discussed below.

Significant channel changes are evident in the reach upstream of Essil (Figure 5.2). Lateral migration of the reach downstream of the road bridge has taken place since 1870, with the channel shifting ca. 150m to the east between 1870 and 1970. This lateral migration has since been checked with extensive bank protection. Downstream, the river in the past occupied the floodplain to the west of the present channel forming a complex of braided channels (Figure 5.2), evidence of which remains today.

Downstream of Essil, the main 1870 and 1903 channels are to the east of the present channel and over the last 120 years the channel has migrated over ca. 400m of floodplain (Figure 5.2). The Essil bend is identifiable in 1971, but with a curve of lower radius than in 1995. The channel configuration just upstream of the viaduct has changed substantially.

Downstream of the viaduct, the planform is, and has been since at least 1870, continually changing as the channel switches and migrates over a ca. 150-200m wide floodplain (Figure 5.2) (Riddell and Fuller 1995; Gemmell et al. 2000). The frequency of changes in this reach render it difficult to identify any long term trend, as changes noted from comparison of maps and aerial photographs do not necessarily represent gradual channel migration, but are the result of a series of changes of varying magnitude and direction. Planform changes occur rapidly during flood events and frequently in this reach and it can be concluded that this reach is highly active and unstable.

Cross-section information is required in order to determine temporal changes in the volumes of sediment stored within the floodplain and in the positions of major channels. Cross-sections of the lower Spey floodplain were determined photogrammetrically at the sections shown in Figure 4.1 (A-G), using 1967 aerial photography and compared to cross sections surveyed at the same locations in 1889 (Lewin and Weir 1977). Comparison of these sections (Figure 5.3) indicate very little difference in the mean level of the sediment surface, despite changes in the location and number of channels present (Lewin and Weir 1977). Sections at similar locations to B and G were resurveyed in this study (see Section 5.2.1).

The historical evidence presented in this section suggests the lower Spey has become increasingly stable with increasing amounts of vegetation over the last 200 years. This is especially pronounced in the upper part of the reach, where the flow is confined to one main channel. The decrease in braiding, decrease in active channel width and increase in channel stability may reflect a decrease in the sediment input to the reach and/or a decrease in the magnitude and frequencies of flooding.

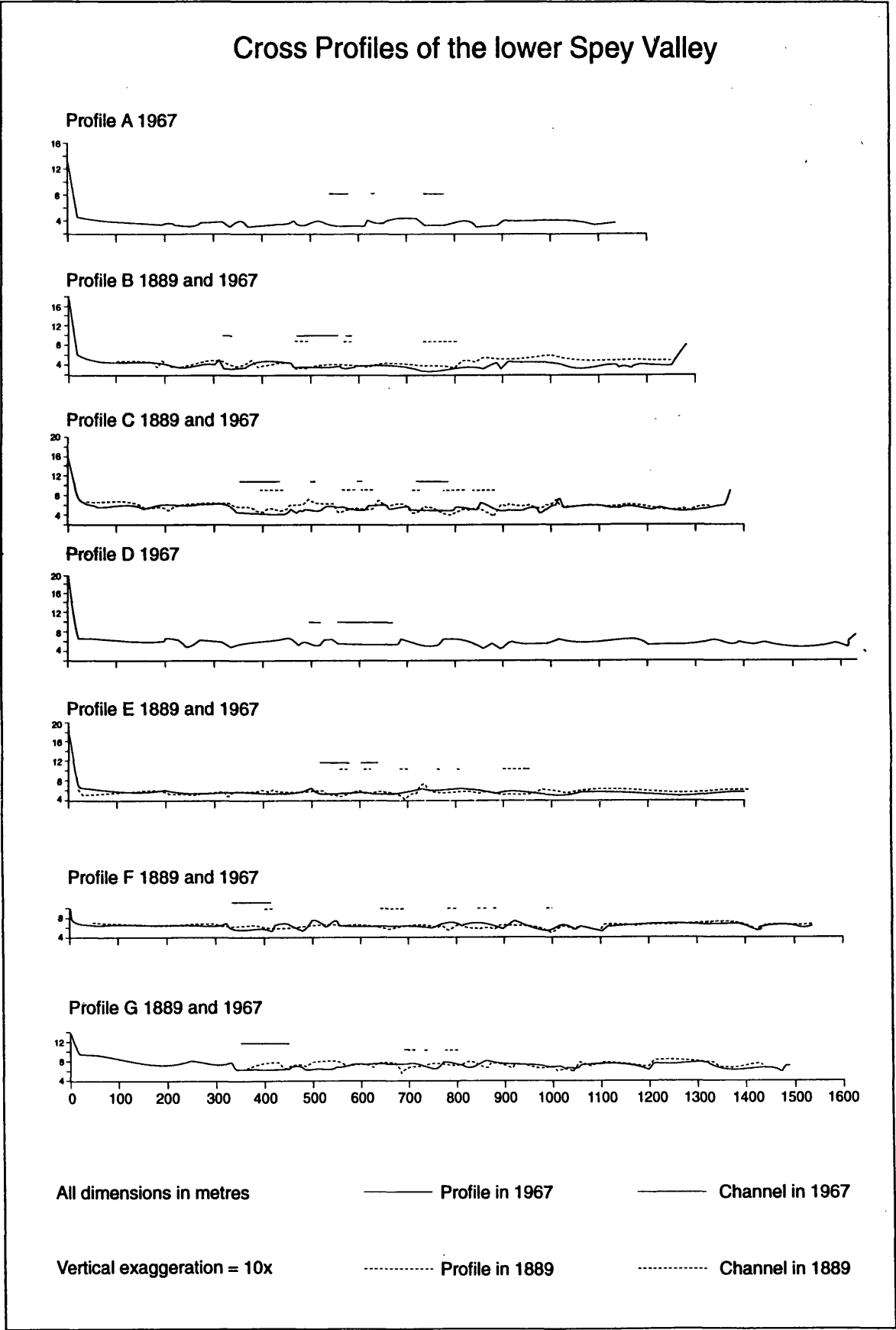


Figure 5.3: Comparison of cross-profiles of the lower River Spey (1889 and 1967) (source: Lewin and Weir 1977). Profiles B and G correspond with profiles 13 and 11, respectively, which were surveyed in 1996, 1997 and 1999 (see Section 5.2.1).

5.2 Short-term fluvial change

This section presents the morphological changes recorded between July 1993 and October 1999. Field mapping and surveying began in October 1995 and prior changes are plotted from aerial photography (Section 4.1.3). Four principal data sources are combined to compute the contemporary sediment budget:

- (1) Cross-section changes quantified from repeat surveys of fixed cross-sections (Section 5.2.1). These are used to estimate sub-reach storage changes and calculate the reach-scale sediment budget using equation 2.3, Section 2.1.1.2. Repeat cross-section surveys are also used to define the depth of mobile sediment, d , (or scour depth) at different locations down the reach. This is defined as the vertical distance from the bank top to the minimum elevation of the channel bed;
- (2) Sediment storage volumes quantified from reach scale surveys of bar morphology (Section 5.2.2). These are used to compute the volume of stored sediment in the reach and to estimate the depth of mobile bar sediment, d . This is calculated as the volume to area ratio of each bar.
- (3) Planimetric area changes quantified from repeat field mapping and air photo analysis (Section 5.2.3). Areas of erosion and deposition were digitised by overlying consecutive maps. The areas were computed in the GIS and converted to volumes by multiplying by the depth of mobile sediment, d , calculated from detailed bar (Section 5.2.2) and cross-section surveys (Section 5.2.1);
- (4) Surface and sub-surface characteristics of the fluvial sediment (Section 5.3).

5.2.1 Cross-section surveys

Bedload transport rates have been calculated from repeat surveys of river cross-sections using the concept of sediment continuity (Section 2.1.1.2). This approach has been applied to rivers over a variety of temporal and spatial scales, with cross-section spacing varying from 1m (Lane et al. 1994; Wathen 1995) up to 2km (Griffiths 1979; McLean 1990) depending on system scale. Typically spacing is ca. 2-3 times the mean channel width (Section 2.1.1.2). The extent to which repeat cross-section surveys, particularly those at wider spacings, can realistically estimate bedload transport is open to question (Section 2.1.1.2) and will be explored herein.

This section presents the results from three repeat cross-section surveys of the lower Spey (average downstream spacing is 325m, ca. 2 in dimensionless form, Figure 4.1). The cross-

section data are used primarily to provide the depths of mobilised sediment for area to volume conversions (Section 5.2.3), but are also used to estimate bedload transport.

Cross-section change

Cross-sections were surveyed on three occasions (Figure 5.4). Section 9 marks the upstream limit of the study reach. Some of the profiles in survey 1 extend across the entire floodplain and these display the typical form of wandering river profiles (e.g. sections 11 and 13). In addition to the present main channel, other channels are present at various locations across the floodplain, representing locations of former main channels. These presently serve as conduits for flood waters (e.g. during the September 1995 flood event water depths exceeded 1m in several of these) and indicate potential routes for new channels to take following avulsion. Surveys 2 and 3 were of the active channel and bars only.

Several sections remained fairly stable between surveys (e.g. sections 9 and 14), whilst others experienced bar erosion (section 10), mid-channel riffle development (sections 12 and 12.5) or major bank erosion (section 15). Changes such as bar and bank erosion can be identified and quantified from repeat planimetric mapping (Section 5.2.3), but other changes, such as mid-channel riffle development or channel scour and fill can only be identified from repeat surveys of the channel bed.

The depth of sediment mobilised in specific erosion events is obtained from repeat cross-section survey. For example, the ca. 50m of left bank retreat recorded at section 15 between surveys 2 and 3 (Figure 5.4) released sediment to depths of 2.7m (bank top to maximum depth of the bed). Mobilisation depths (*d*) of bank sediment for specific locations are computed from the cross-section data (Table 5.2) and applied in Section 5.2.3 to convert area changes to volumes. There is no downstream trend in the depth of mobile sediment (Table 5.2) and if no surveyed depth is available at a particular location the mean bank-full channel depth (2.631m) was used.

Cross-section	bank	depth (<i>d</i>) / m
10	left	2.761
12	left	2.212
12.5	left	2.738
12.5	right	2.539
13	left	2.515
14	right	2.949
15	left	2.703
mean		2.631 ± 0.177

Table 5.2: Depths of sediment mobilised in specific bank erosion events. These depths are applied to convert areas of erosion to volumes (Section 5.2.3).

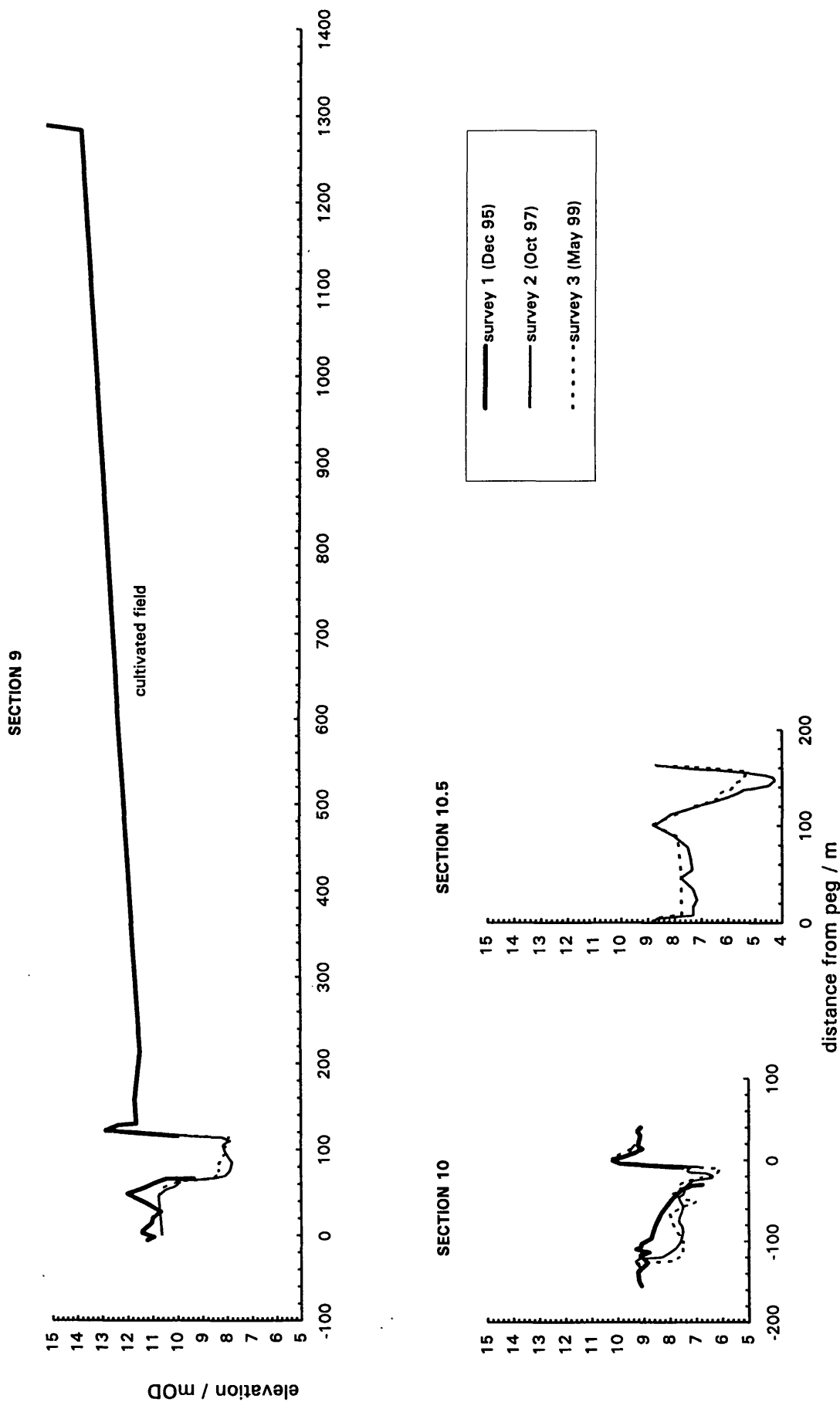


Figure 5.4: Lower River Spey cross-sections

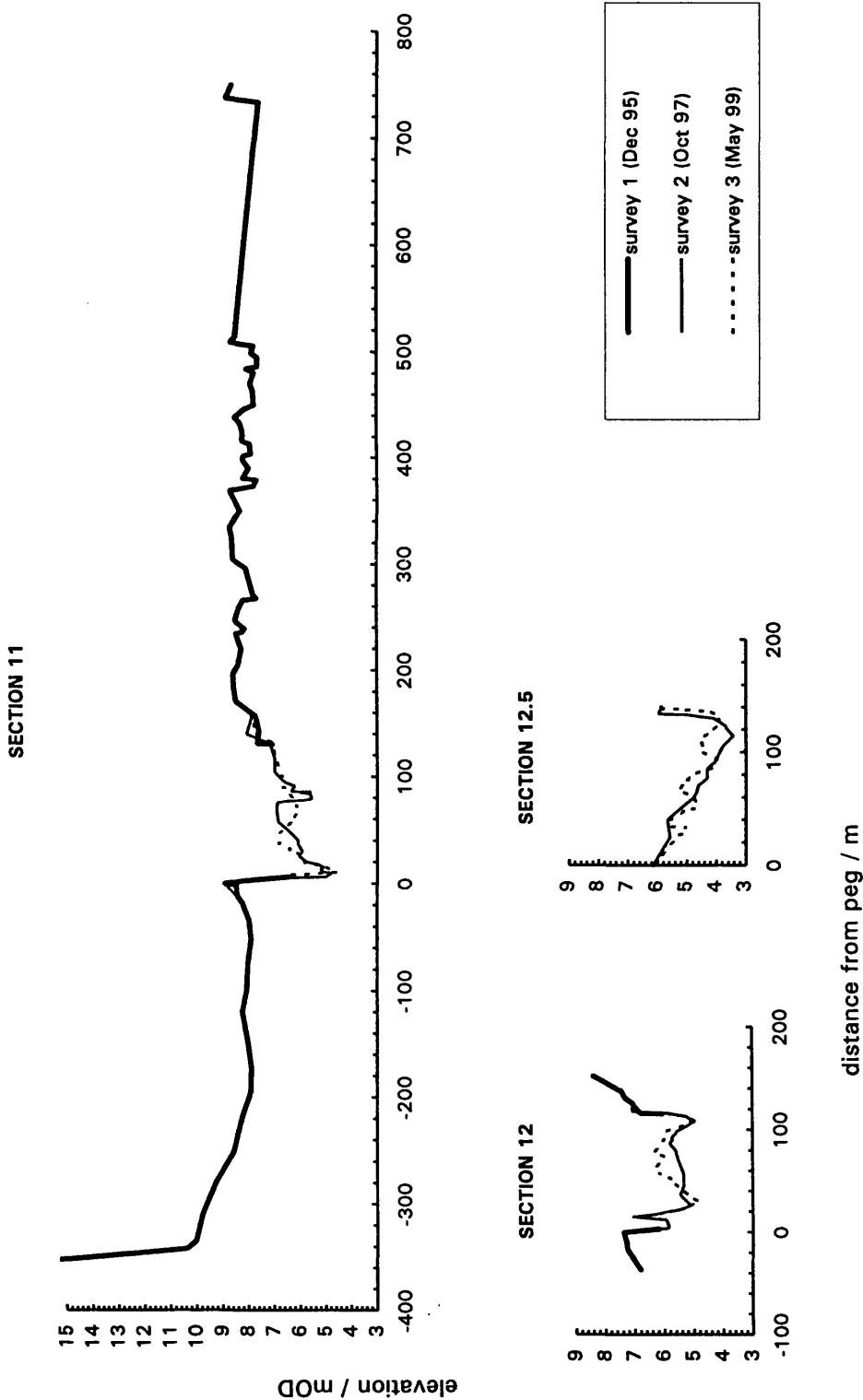


Figure 5.4: Lower River Spey cross-sections

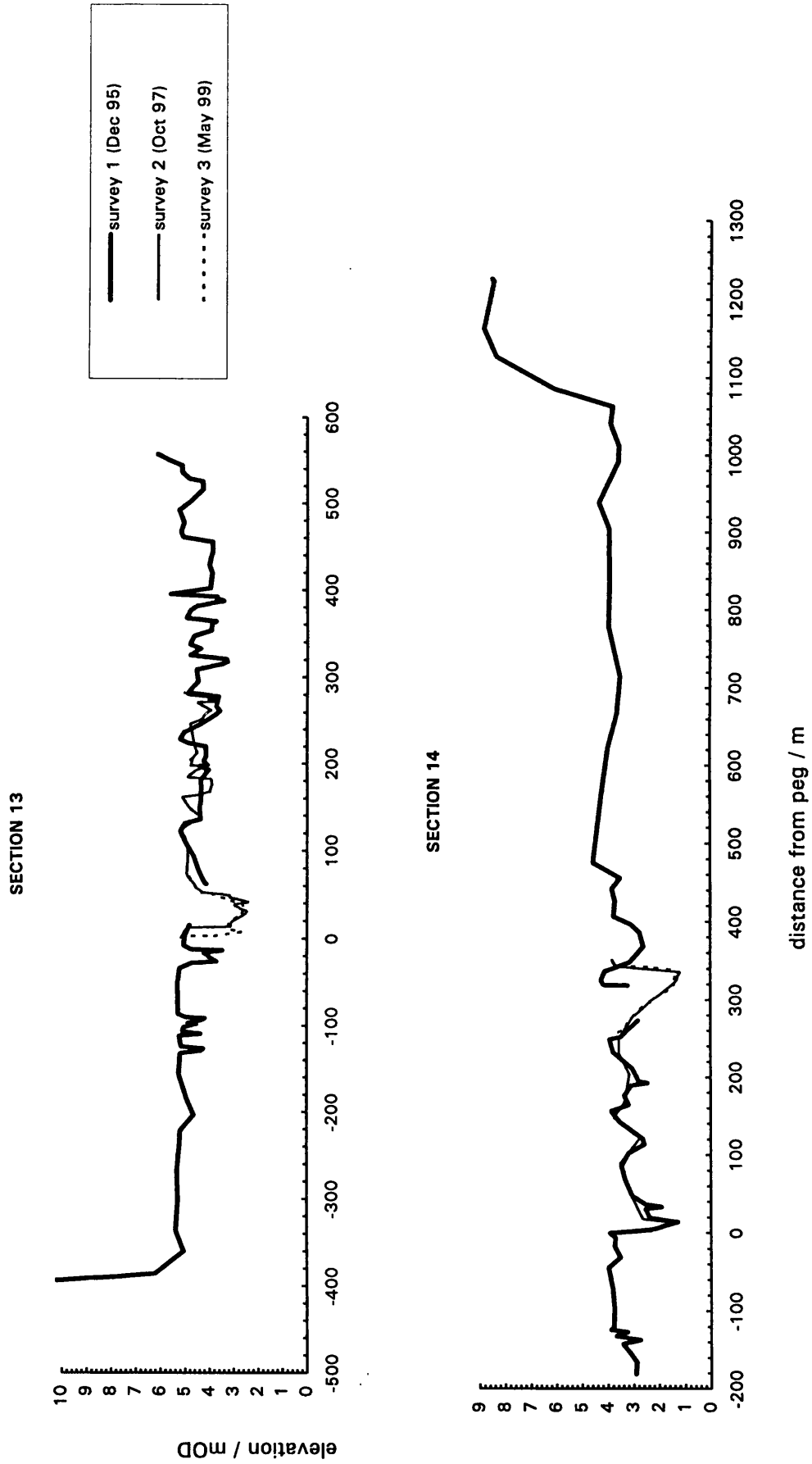


Figure 5.4: Lower River Spey cross-sections

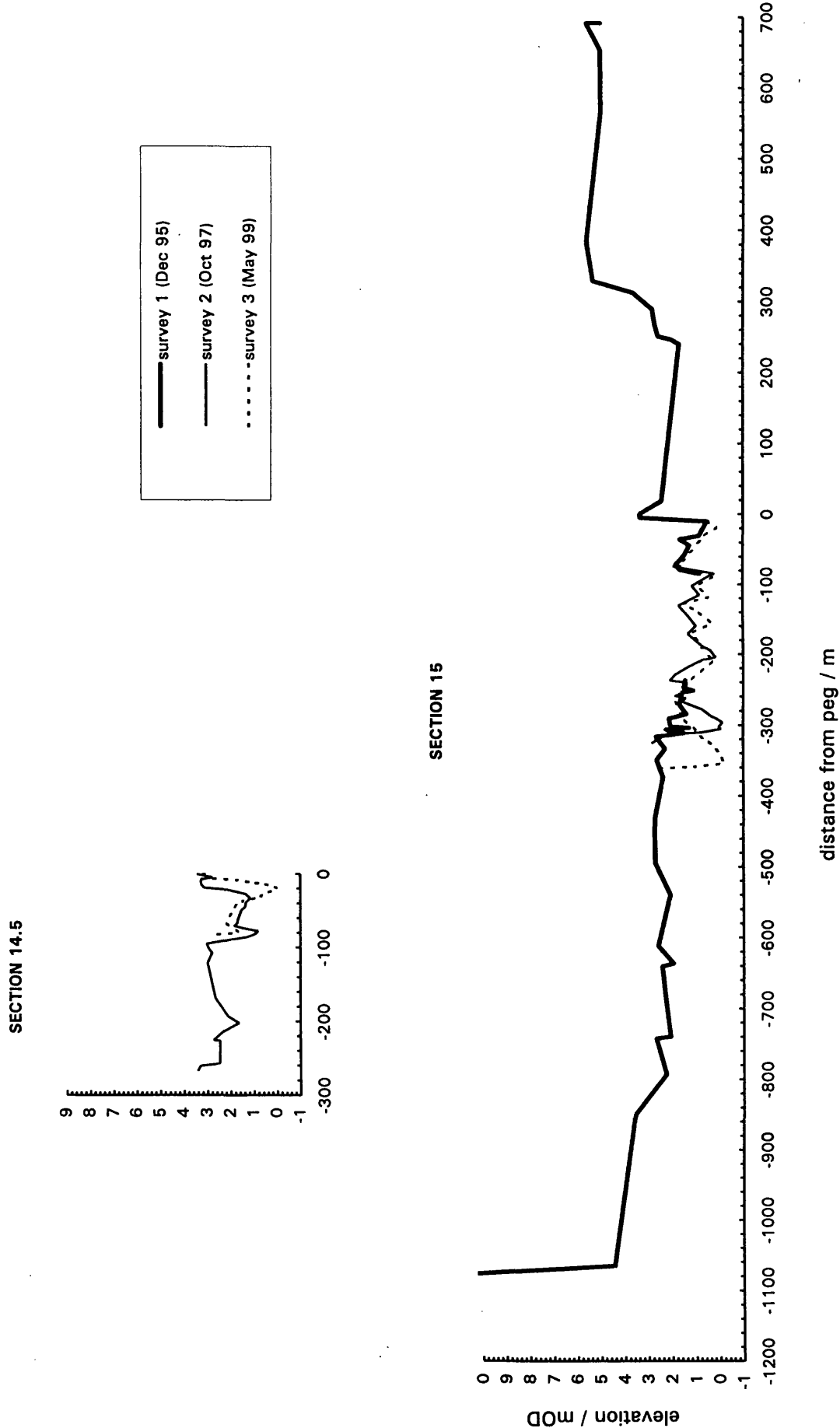


Figure 5.4: Lower River Spey cross-sections

To compute bedload transport rates using cross-sections, the net change in area between surveys at each section is converted to reach volume change by multiplying by the representative reach length (Section 2.1.1.2). Calculating area change between surveys 1 and 2 is problematic as the wetted perimeter was not surveyed during survey 1 (Figure 5.4). However, where possible, the channel dimensions were estimated and reconstructed using survey 2 data. Net volume changes are calculated for the sub-reach between each section (Table 5.3) assuming that the change in area at a section is representative over the distance between it and a half distance to each adjacent section (equation 2.5, Section 2.1.1.2).

Section no.	Net area change / m ²		Sub-reach no.	Sub-reach Length / m	Volume Change / m ³
	Dec 95 to Oct 97	Oct 97 to May 99			
9		+8.50			
			1	747	-75
10	-74.87	-8.70			
			2	157	+3 069
10.5		+47.71			
			3	242	+5 736
11		-0.36			
			4	314	+4 136
12		+26.72			
			5	311	+5 319
12.5		+7.52			
			6	316	-1 067
13	+47.85	-14.27			
			7	301	-530
13.5		+10.75			
			8	246	+1 192
14	-5.02	-1.05			
			9	208	-2 099
14.5		-19.10			
			10	413	-29 234
15		-122.61			

Table 5.3: Net area change (m²) at cross-sections. These are converted to volumetric changes (m³) for each reach for the period Oct 97 to May 99.

Net area changes at cross-sections range from +47.71 m² at section 10.5 to -122.61 m² at section 15 (Table 5.3). When integrated between adjacent sections, these area changes represent large volumetric changes (e.g. a net loss of 29 234 m³ of sediment from sub-reach 10 between Oct 97 and May 99). These changes will be compared to sub-reach storage changes quantified using the planimetric technique (Section 5.2.3).

5.2.2 Reach-scale surveys of storage zones

At the outset of this study, it was intended that repeat surveys of bar morphology would be used to quantify volumetric changes in each sub-reach to compute the sediment budget (cf. Lane et al. 1995, Section 2.1.1.2). However, due to the lack of major morphological change

during the study period (Sections 4.1 and 5.2.3) the bar surveys are also used to quantify the volumes of sediment stored in bars and islands and to give values for the depths of sediment (d) involved in the planimetric changes identified in Section 5.2.3.

Digital elevation models (DEMs) of each bar were constructed as outlined in Section 4.6.1.1. Each bar was delimited by the surveyed perimeter of the waters edge (e.g. Figure 4.3) and storage volumes were computed using planar, sloping, base levels defined by a regression surface of the lowest level of the bed at surveyed cross-sections in the reach (Section 4.6.1.4). With a mean survey point spacing of less than 6.5m, bar volumes (Table 5.4) are considered to represent the actual storage volumes to within $\pm 2.5\%$ (Section 4.6.1.6).

bar no.	date of survey	volume m^3	area m^2	volume / area (d) m	description
1	Jun-97	1 597	1 180	1.35	bare gravel
2	Jun-97	8 490	5 397	1.57	bare gravel
3	Jun-97	577	311	1.86	bare gravel
4	Jun-97	7 683	4 877	1.58	bare gravel
5	Jun-97	709	514	1.38	bare gravel
6	Jun-97	2 459	1 532	1.61	bare gravel
7	Jun-97	3 210	1 882	1.71	bare gravel
b (1)	Jun-97	94 565	52 630	1.80	vegetated
b (2)	Jul-97	87 106	37 624	2.32	vegetated
b (3)	Jul-97	5 049	3 701	1.36	bare gravel
b (whole)		187 772	94 196	1.99	vegetated
8	Jul-97	107 073	51 293	2.09	vegetated
9	Jul-97	465	291	1.60	bare gravel
10	Jul-97	25 561	10 694	2.39	bare gravel
11	Jul-97	60 459	35 320	1.71	bare gravel
riffle	Jul-97	11 493	6 713	1.71	bare gravel
12	Jul-98	44 192	30 060	1.47	bare gravel
13	Jul-97	19 197	12 475	1.54	bare gravel
14L	Jul-98	8 466	5 136	1.65	bare gravel
14H	Jul-98	10 394	5 300	1.96	bare gravel

Table 5.4: Lower River Spey bar volumes and areas. Locations are given in Figure 5.5.

A detailed survey of the bars and islands upstream of section 11 was deemed unnecessary because most of these bars/islands were well vegetated and relatively stable. However, all bars within the active channel zone downstream of section 11 were surveyed (Figure 5.5). In the field it is difficult to accurately delimit active (or potentially active) bars as they often merge into higher, more stable deposits (e.g. bar 10). The larger bars (e.g. bars 8 and b) are partially vegetated with the vegetated area increasing during the study, indicating the lack of activity and increasing stability of bar surfaces. From this analysis, the total volume

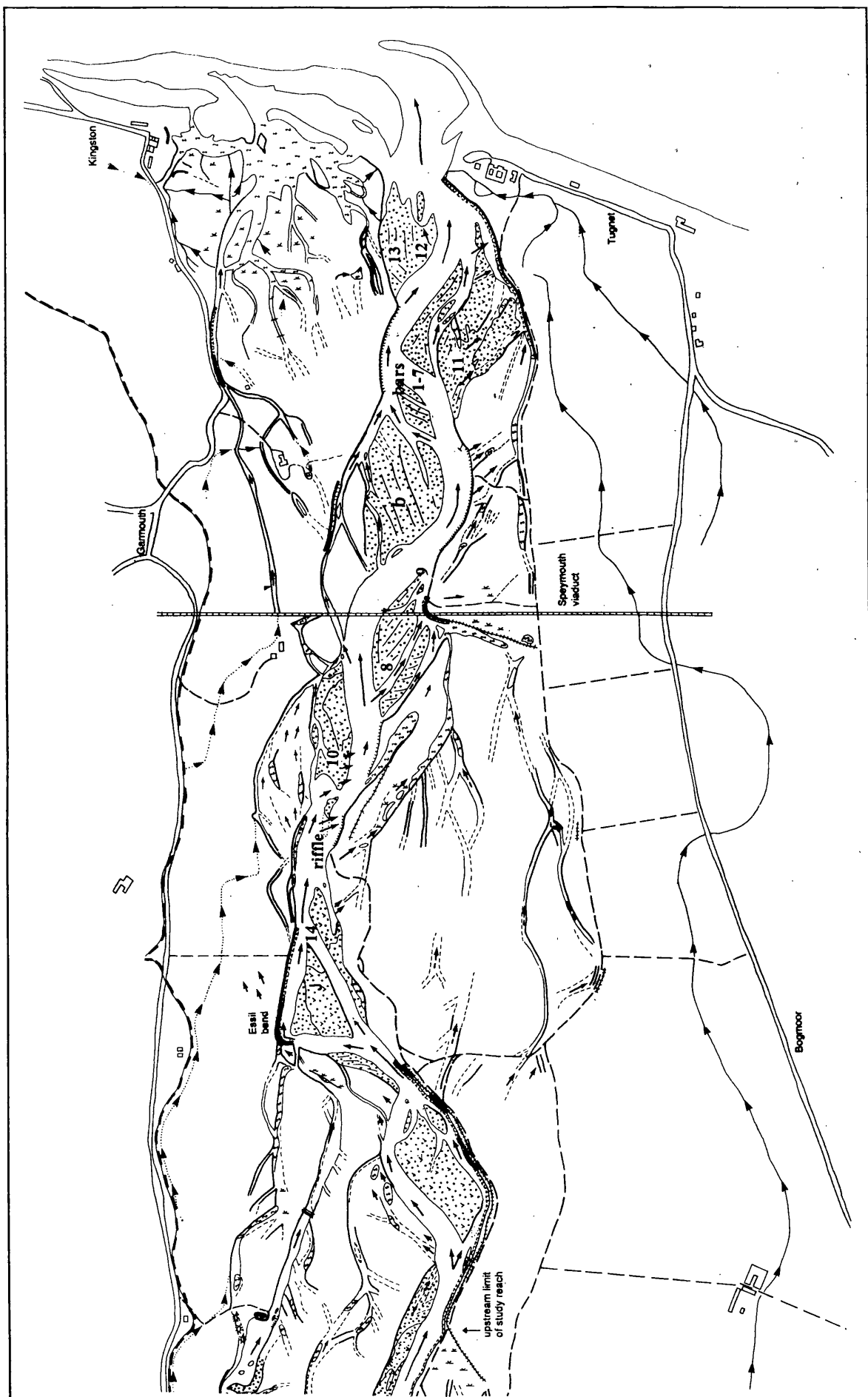


Figure 5.5: Location of bars surveyed on the lower River Spey

of sediment stored in active (or potentially active) bars downstream of section 11 is 475 000 m³.

Bar volumes range from 579 m³ (bar 1) to 187 772 m³ (bar b) (Table 5.4). Bars b and 8 are the largest stores of sediment, with a combined volume of ca. 295 000 m³. This represents almost 40 times the predicted 8 000 m³a⁻¹ supply of sediment from the River Spey to the coast (Dobbie & Partners 1990). A major flood and channel avulsion could potentially re-activate this sediment producing a large sediment input to the coast.

Volume to area ratios (Table 5.4) indicate the relative elevation and stability of the deposit, and represent the depths of mobile sediment (d) for each bar, assuming that the sloping datum represents a maximum scour depth. Ratios vary from 2.39m to 1.36m (mean = 1.73m, standard deviation = 0.29m). The higher ratios are typically the higher, more stable, vegetated or partially vegetated, deposits (e.g. bar 8) while lower ratios tend to be lower, more active bars (e.g. bar 1) or emerging riffles. Bar 10 has the highest ratio, which may reflect the higher vegetated terrace forming the left boundary of the active bar (Figure 5.5).

The ratios in Table 5.4 were used to convert area changes identified in Section 5.2.3 into representative volumes. If no field survey data was available for a particular location, the mean volume/area ratios were used (1.653 and 2.048 for unvegetated and vegetated bars and islands, respectively). This differs from other studies where the depth of mobile bar sediment is estimated by the elevation difference between the bar top and thalweg (Ham 1996). This is likely to over-estimate volumes of erosion and deposition, as the bars here rarely maintain a uniform elevation relative to the waters edge. The method used herein is considered more representative of the volume changes that are actually occurring as they reflect the topography of individual bars and islands, which varies with their hydraulic location (Church and Jones 1982) and depositional history.

5.2.3 Planimetric changes

The record of channel change on the lower River Spey is based on field mapping and aerial photography (Table 5.5). Three geomorphological maps were compiled in the field (December 1995, December 1997 and April 1999). Large channel changes, which occurred prior to the start of this study, are quantified by mapping the channel planform from aerial photography taken in July 1993 and July 1994. The maps were rectified, digitised and input into a GIS to assess changes over time (Section 4.1.3). Areas of erosion and deposition were digitised by overlying consecutive maps in the GIS and converted to volumes by multiplying by the relevant depth of mobile sediment, d (Tables 5.2 and 5.4).

date	source	approx. discharge at time of photography / mapping ($\text{m}^3 \text{s}^{-1}$)	channel width at section 14 (from rating curve) (m)
13 July 93	1: 10 000 aerial photography	32	40.6
mid July 94	1: 5 000 aerial photography	24	37.3
Oct - Dec 95	1: 5 000 field mapping	90	61.3
1-2 Dec 97	1: 10 000 field mapping	47	48.1
29-30 Apr 99	1: 10 000 field mapping	52	49.8

Table 5.5 : Record of planimetric channel change at the lower River Spey.

Ideally, for accurate quantification of surface bar and bank area changes, all photography and mapping should be carried out at similar discharges. At higher discharge the water surface width is increased, thus decreasing the surface exposure of bars and islands. Stage changes are likely to have a greater impact on wider, shallower, braided reaches (e.g. the most downstream reach of the lower Spey). The mean daily discharge during field mapping and photography varied from $24 \text{m}^3 \text{s}^{-1}$ to $90 \text{m}^3 \text{s}^{-1}$ (Table 5.5), with the maximum occurring during initial mapping in October and December 95. The initial map was based on 1994 aerial photography (Section 4.1.3) and the presence of submerged bars and riffles were noted during mapping. If bars and riffles identified in the 1994 photographs were present as submerged features in 1995, any perceived area change between dates was discounted from the sediment budget calculations. In the shallower more braided reach, downstream of section 12.5, the effect of stage on the surface exposure of bars was less easy to assess in the field and a correction was applied as follows.

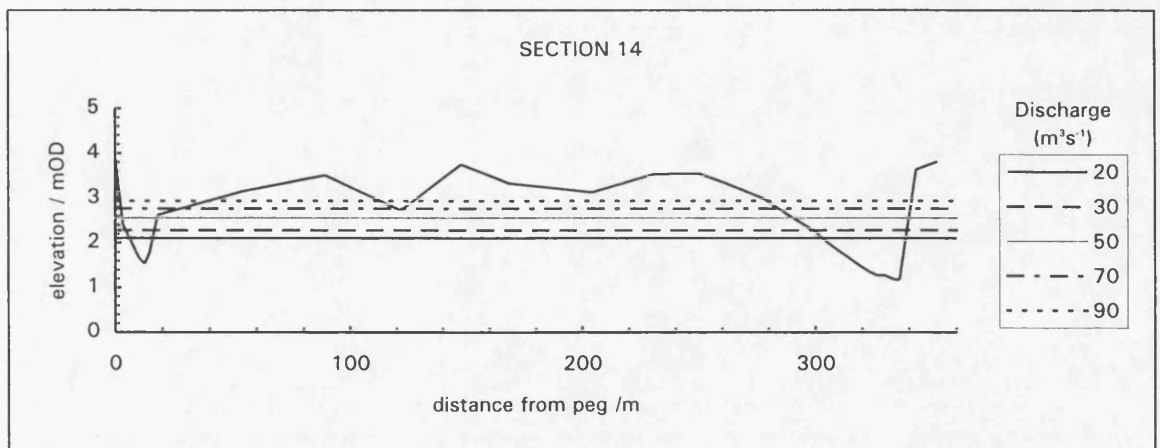


Figure 5.6: Variation in water surface elevation and channel width with discharge at section 14. Note the effect on the surface exposure of the left bank bar.

Water surface elevations and channel widths were calculated for each discharge (Table 5.5 and Figure 5.6) using the rating curve derived for section 14 (see Section 5.4.1). On the lower reach of the Spey, there is minimal change in channel width and hence the surface

exposure of bars for discharges up to $50 \text{ m}^3 \text{ s}^{-1}$ (Figure 5.6) and so no correction is applied for the majority of flows observed. However, for a discharge of $90 \text{ m}^3 \text{ s}^{-1}$ the surface bar exposure decreases substantially (Figure 5.6 and Table 5.5) so a correction is applied to the volume changes measured when comparing the Dec 1995 map. Volume change (ΔV) is calculated using:

$$\Delta V = \Delta \hat{V} \pm V_{\text{corr}} \quad (5.1)$$

where $\Delta \hat{V}$ is the ‘measured’ volume change and the correction factor, V_{corr} , is estimated as:

$$V_{\text{corr}} = \delta w \cdot L \cdot d \quad (5.2)$$

where L and d define the length and bed material depth of the bar/island, respectively and δw is the change in channel width between mapping periods. For an increase in stage between dates ($\delta w > 0$), the corrected volume of erosion (ΔV) is equal to the ‘measured’ volume ($\Delta \hat{V}$) minus the apparent erosion volume caused by the stage change (V_{corr}). Deposition volumes are corrected by adding V_{corr} to the measured volume of bar deposition. The same rules apply for a decrease in stage between dates as $\delta w < 0$, so $V_{\text{corr}} < 0$. The correction was applied to bars adjacent to the main channel downstream of section 12.5.

Morphological changes between consecutive maps are discussed below. $280 \text{ m}^3 \text{ s}^{-1}$ is considered the threshold flow for initiation of morphological change (see Section 5.4.3).

July 1993 - July 1994 (Peak discharge = $460 \text{ m}^3 \text{ s}^{-1}$, number of events exceeding $280 \text{ m}^3 \text{ s}^{-1}$ = 11)

Major channel changes occurred in this period (Figure 5.7). In the upstream part of the study reach the outer meander bends were erosional and the inner bends depositional, with up to 20m of bank erosion at Essil (Figure 5.7). Downstream, there was an avulsion, with the 1994 channel flowing ca. 110m west of its 1993 course at the Speymouth viaduct, eroding a wooded island. Local sources report that this channel change occurred during a flood event in January 1994 (Stratton 1996, pers. comm.). This avulsion created major changes in the downstream reach, north of the viaduct (Figure 5.7). Large areas of bare gravel became exposed in the location of the former 1993 channel, while the new channel eroded vegetated and unvegetated gravel deposits, providing access to substantial volumes of sediment for potential transport to the coast.

July 1994 - December 1995 (Peak discharge = $718 \text{ m}^3 \text{ s}^{-1}$, number of events exceeding $280 \text{ m}^3 \text{ s}^{-1} = 7$)

Between July 1994 and December 1995, up to 20m of erosion occurred on the outer bank at the apex of Essil bend (Figure 5.8), resulting from the flood of 10-12th September 1995. The bank was over-topped, leading to flooding and erosion at Essil which prompted emergency bank protection works undertaken in October 1995 (see Gemmell et al. 2000). Pertinent to this study is the dredging of the eastern channel, which was occupied only during high flow conditions. A ca. 30m wide, 0.5m deep channel was dredged to allow 150m of rip-rap bank protection to be put in place on the west bank of the western channel. The excavated gravel was not removed from the river, and was used either to infill the eroded west bank, block off the western channel or was dumped on the adjacent island. By December 1995, the artificially deepened eastern channel continued to convey Spey waters, even at moderate and low flows, and fluvial action further widened and deepened the channel (Figure 5.8). These changes have implications for the supply of sediment to downstream reaches.

Bank erosion continued at the Speymouth viaduct, with a further 50m recorded in the four months prior to October 1995 (Figure 5.8). Deposition was observed on the inner bar. However, due to differences in river stage (Table 5.5) the area of deposition shown on the map is likely to be an underestimate. Similarly, areas of bar erosion in this reach are likely to be overestimated. Notwithstanding the above reservations, significant bank erosion was recorded on the east bank at section 14 (up to 20m) and on the west bank, just upstream of section 15 (up to 80m). This major channel change eroded an entire fairway from the Kingston and Garmouth golf course (Plate 5.1), releasing large volumes of sediment.

December 1995 - December 1997 (Peak discharge = $705 \text{ m}^3 \text{ s}^{-1}$, number of events exceeding $280 \text{ m}^3 \text{ s}^{-1} = 9$)

The July 1997 flood, with a peak discharge of $705 \text{ m}^3 \text{ s}^{-1}$ and a return period of ca. 3 years (Section 3.3.1), created the largest morphological changes in the lower River Spey during the field study period (Figure 5.9). The entrance to the western channel at Essil was completely infilled with sediment between December 1995 and December 1997 (Figure 5.9). This occurred gradually, although a large volume was deposited during the flood. Quantification of the volume of this deposit gives the upstream boundary condition of volumetric transport rate into the reach (Q_i) necessary to compute the reach scale sediment budget (equation 2.3, Section 2.1.1.2). Other changes in this reach include erosion of the

Figure 5.7: Planimetric channel change (July 93 - July 94)

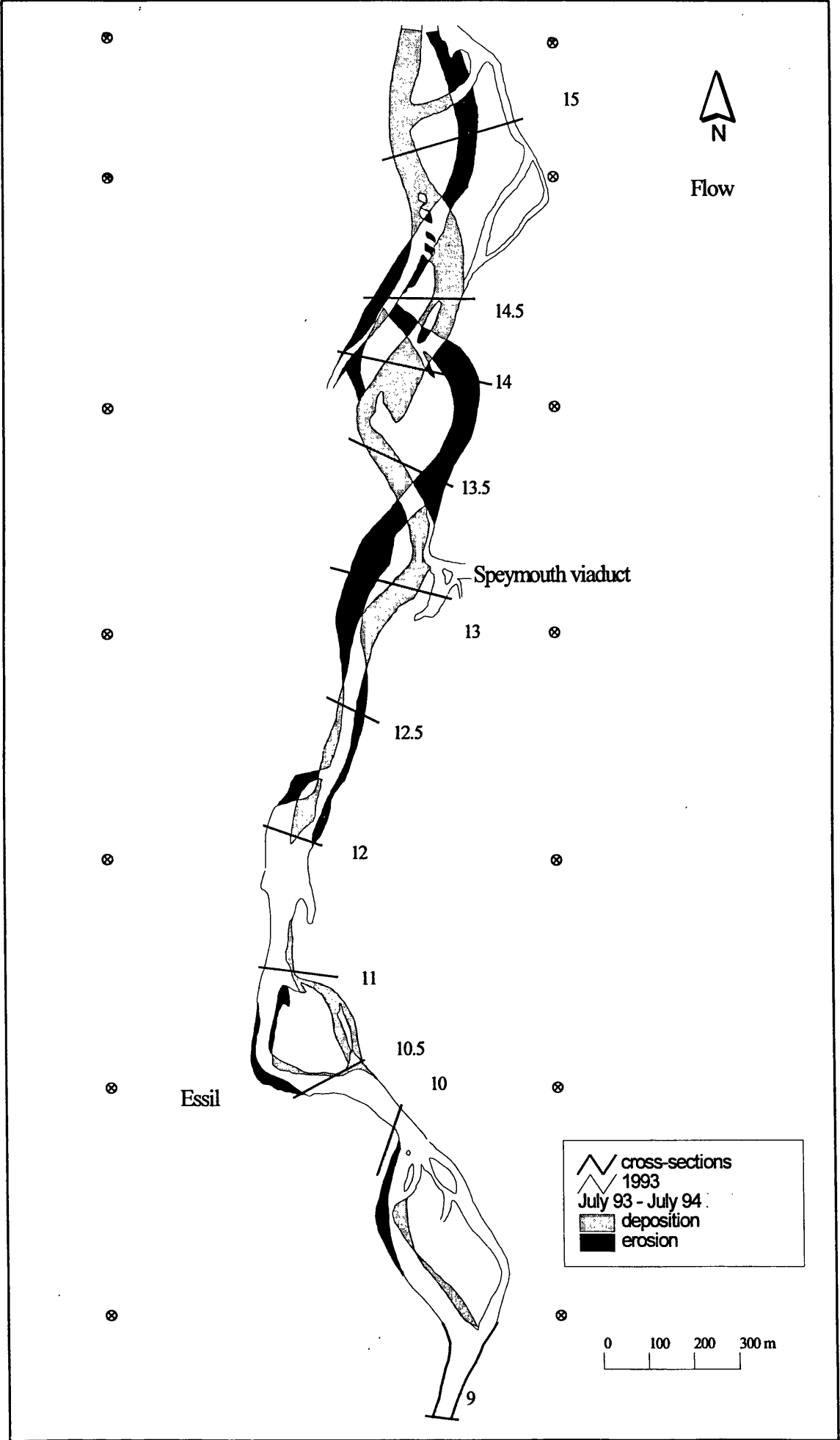


Figure 5.8: Planimetric channel change (July 1994 - December 1995)

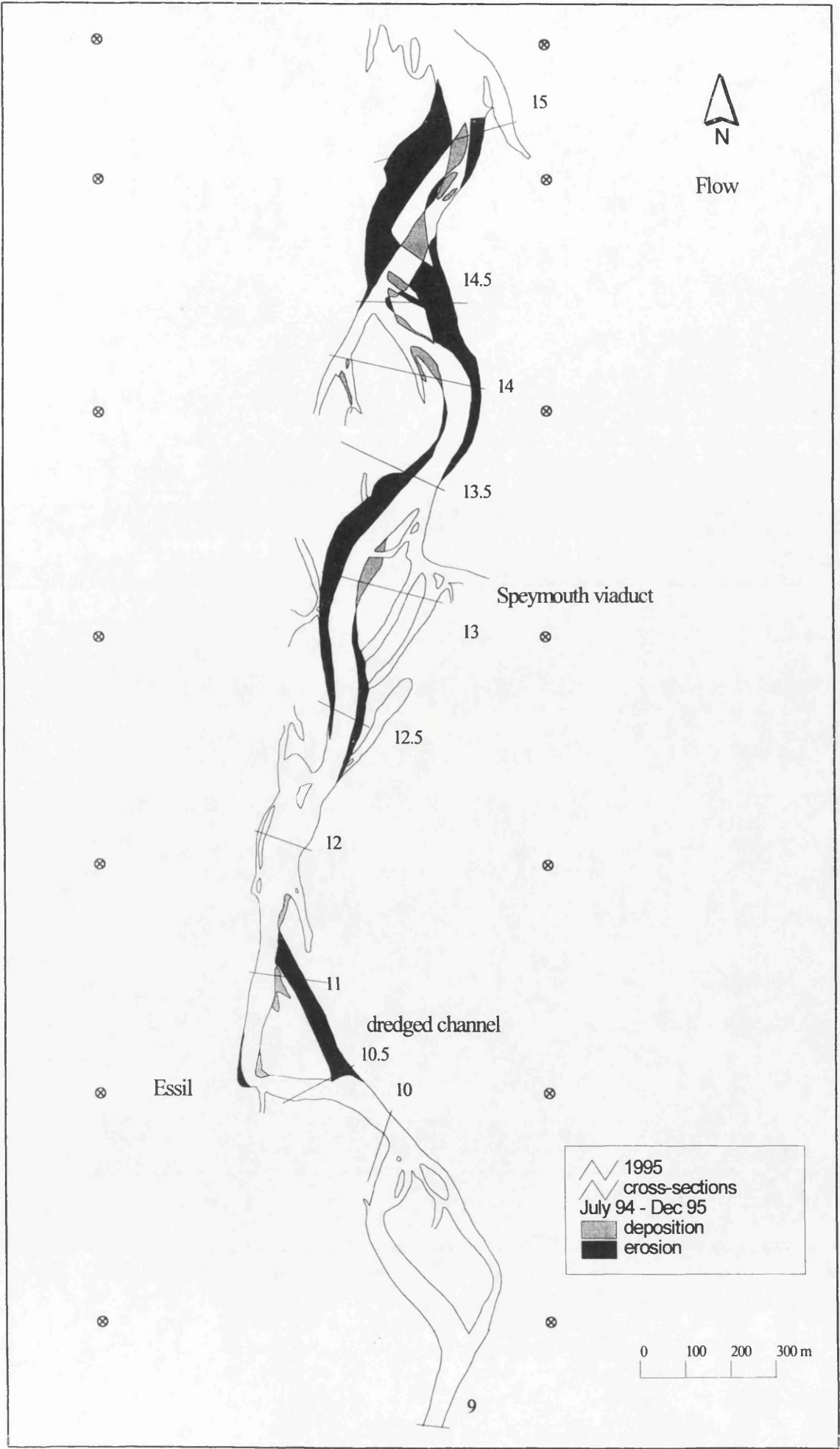


Figure 5.9: Planimetric channel change (December 1995 - December 1997)

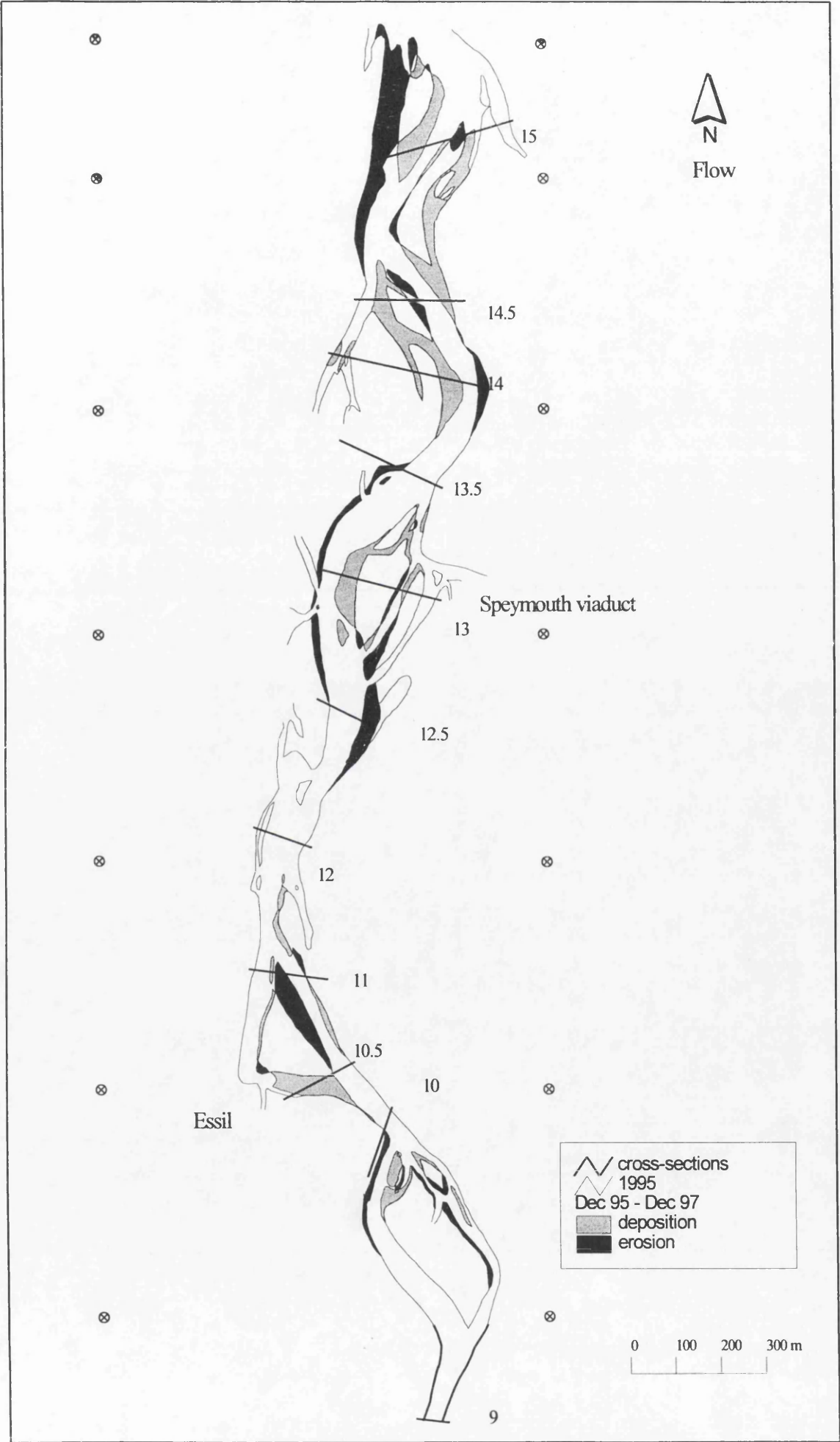
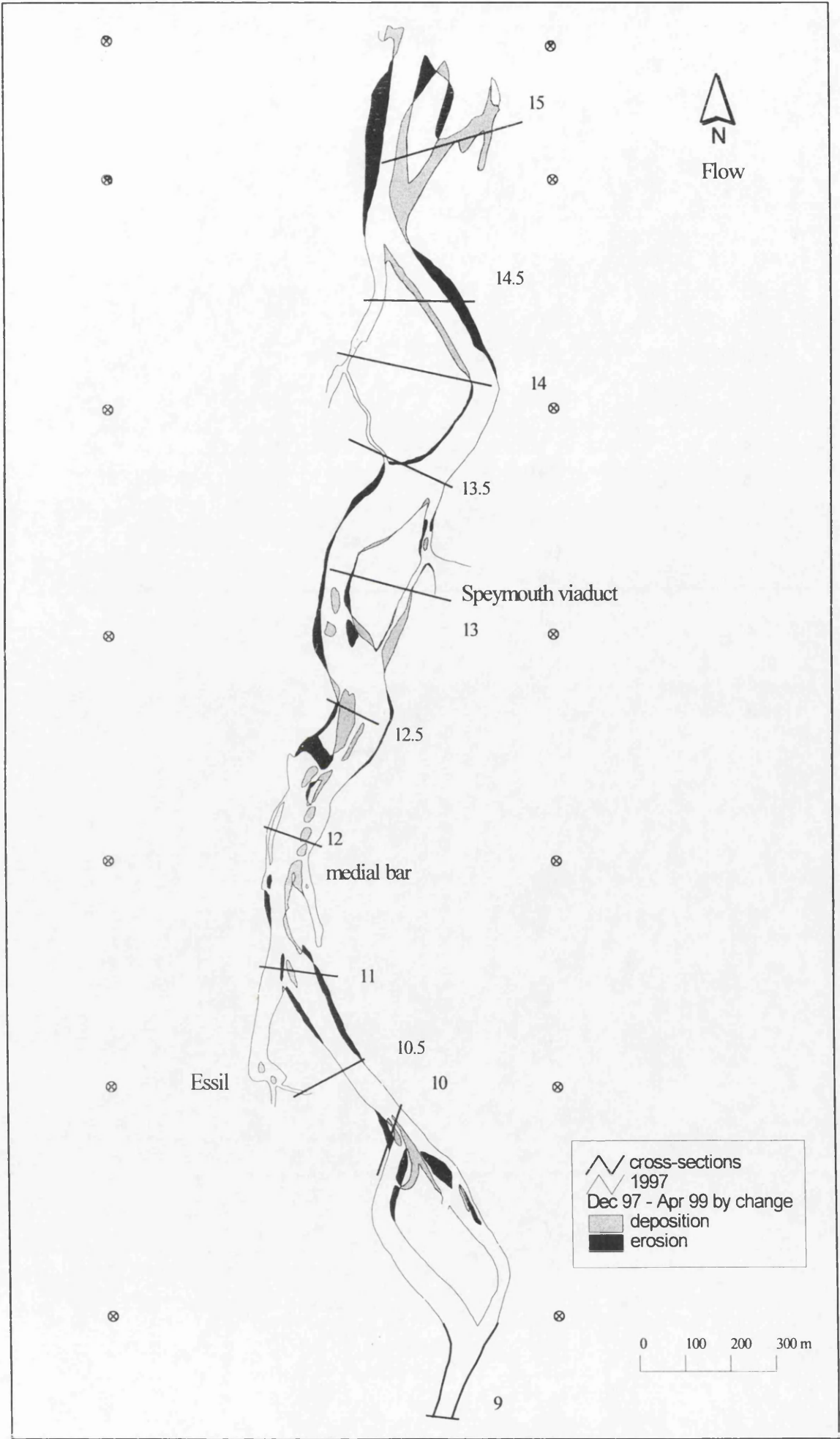


Figure 5.10: Planimetric channel change (December 1997 - April 1999)



bar immediately downstream, which now forms the main west bank of the river and deposition of a low bar on the east bank (Figure 5.9).

Downstream, the main changes between December 1995 and December 1997 were enhanced erosion on the outer banks of meander bends, particularly the east bank at section 12.5 (ca. 35m), the west bank at the viaduct (ca. 10m), the east bank downstream of the viaduct at section 14 (ca. 25m) and the west bank at the golf course (ca. 30m). The main channel avulsed during the July 1997 flood to cut a new, more direct, course to the mouth through the large west bank bar downstream of section 15 (Figure 5.9).

December 1997 - April 1999 (Peak discharge = $363 \text{ m}^3 \text{ s}^{-1}$, number of events exceeding $280 \text{ m}^3 \text{ s}^{-1} = 3$)

Minor readjustments in planform were recorded in April 1999 (Figure 5.10). Further erosion of the outer meander bends was observed, together with deposition on the inner bends. The emergence of a large bar downstream of Essil was noted: in April 1999 this bar almost continuously connected the upstream right bank bar to the downstream left bank bar (Figure 5.10). Several channels dissect the bar, conveying the main flow from left to right. This dissected bar was a low submerged riffle during previous field surveys (Figure 5.9) and appears to have been a locus for deposition throughout the study. The emergence of this bar unit is not an effect of stage, as the discharge during both field surveys was approximately the same (Table 5.5).

Up to ca. 25m of erosion was recorded on the right bank downstream of section 14, while deposition occurred on the inner bank. Erosion continued on the left bank downstream of the golf course, with a further ca. 40m of recession recorded at section 15 (Figure 5.10). Deposition was observed on the opposite inner bar.

Quantification of Net Storage changes

This section presents the net storage changes (ΔS) in each sub-reach (between two consecutive cross-sections) of the river. Sub-reach 1 is the furthest upstream. The net change in storage volume for a given sub-reach is given by:

$$\Delta S_i = Vd_i - Ve_i \quad (5.3)$$

where Vd_i and Ve_i are the total volumes of deposition and erosion (corrected for stage changes if applicable) in reach i , respectively.

Net storage changes are presented in Appendix B and summarised in Figure 5.11. Storage changes are given as annual bulk volumes ($\text{m}^3 \text{ a}^{-1}$) in Figure 5.11 and mineral volumes (i.e.

corrected for porosity) in Appendix B. Reaches downstream of section 12.5 (sub-reach 6) exhibit larger net volume changes, reflecting the increase in activity and availability of mobile sediment moving downstream. Most reaches experience net sediment loss over the period of analysis. The exceptions are reach 4, which experienced net aggradation throughout the study, due to the emergence of a medial bar (Figure 5.10) and reach 2, which underwent high amounts of aggradation between December 1995 and December 1997, due to the infill of the left-bank channel at Essil. The largest net storage changes occurred between July 1993 and July 1994 and were particularly marked in the sub-reaches downstream of the viaduct, with a change of ca. $-17\,000\text{ m}^3$ in reach 8 (Figure 5.11).

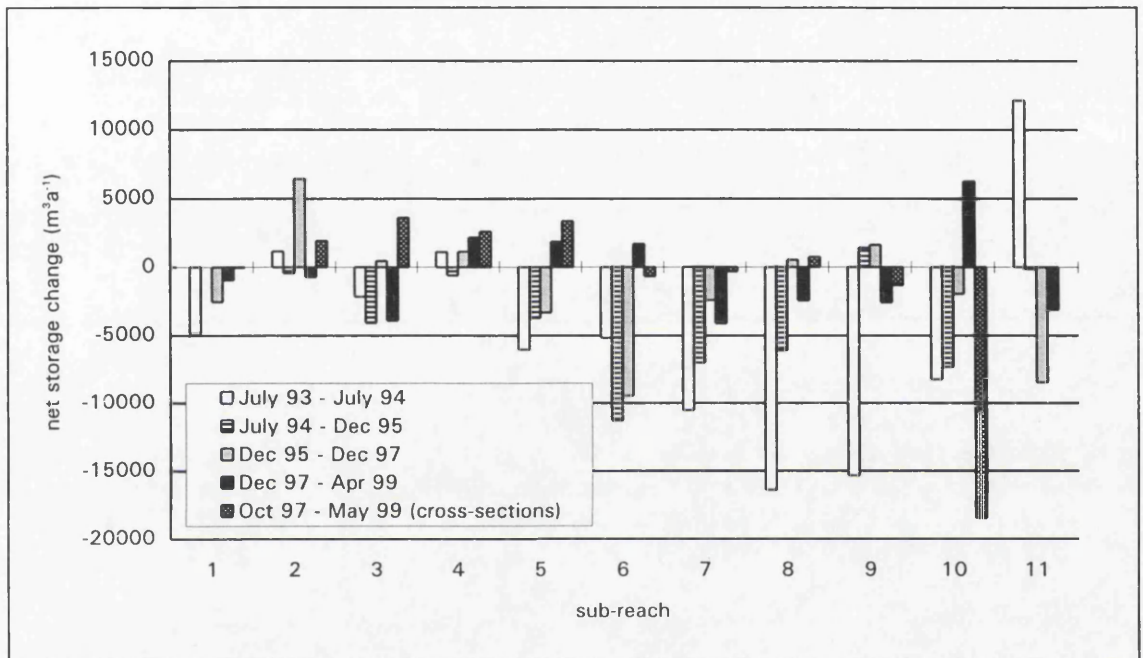


Figure 5.11: Net storage changes ($\text{m}^3\text{ a}^{-1}$) in each sub-reach of the lower River Spey. The sub-reach storage changes quantified from repeat cross-section surveys (Table 5.3) are included for comparison.

This method quantifies lateral erosion and deposition of bars only and does not account for any vertical changes that may have occurred on pre-existing bars. For this reason, all erosion and deposition volumes should be considered to be minimum estimates. The sum of erosion and deposition in each period provides a relative comparison of channel activity (Table 5.6). In each period volumes of erosion exceed volumes of deposition, hence the degradational nature of the channel. The period from July 1993-July 1994 has the largest turnover of sediment, with erosion exceeding $200\,000\text{ m}^3\text{ a}^{-1}$. Total volumes of erosion and deposition decrease by almost 3-fold in the latter two periods, indicating an increased channel stability. The reasons for this will be investigated later.

	Jul 93- Jul 94	Jul 94-Dec 95	Dec 95-Dec 97	Dec 97-Apr 99
erosion ($\text{m}^3 \text{a}^{-1}$)	209 000	94 000	69 000	63 000
deposition ($\text{m}^3 \text{a}^{-1}$)	155 000	54 000	51 000	57 000

Table 5.6: Annual erosion and deposition expressed as bulk volumes (all reaches)

Net storage changes calculated from repeat cross-section surveys for the period October 1997 to May 1999 (Table 5.3, Section 5.2.1) are converted to annual changes and plotted in Figure 5.11. In some reaches (e.g. reaches 4 and 5) these changes are comparable to those calculated using planimetric data between December 1997-April 1999. However, in the majority of reaches the volumes quantified using repeat cross-sections are not only vastly different in magnitude, but also differ in direction from those computed using planimetric mapping techniques (Figure 5.11). For example, reach 10 (between sections 14.5 and 15) has aggradation of ca. $+6\,000 \text{ m}^3 \text{a}^{-1}$ using planimetric mapping, while ca. $18\,000 \text{ m}^3 \text{a}^{-1}$ of net degradation is calculated using repeat cross-section surveys. These differences may be due to section location omitting important zones of deposition or erosion. For example, the east bank deposition in reach 10 is not fully captured by either the upstream or downstream section (Figure 5.10). The assumption that the area change at each section is representative over the distance between it and each adjacent section is problematic at these particular sections (Figure 5.10).

The poor performance of repeat cross-section surveys suggests that in wide, wandering gravel-bed rivers such as the Spey planimetric methods are more reliable, unless a very dense network of cross-section are repeatedly surveyed. Cross-section spacing of 2-3 channel widths is inappropriate on a river of this scale. As a result, storage change estimates from the cross-section data are not considered sufficiently robust to compute the reach-scale sediment budget for the lower Spey. Instead, net storage change estimates from planimetric mapping (Appendix B) are used to compute sediment budgets and transport rates for each period.

5.3 Sediment characteristics

The storage change estimates presented in Section 5.2 are bulk volumes. In order to compute the reach-scale sediment budget and transport rates, storage volumes have to be converted to mineral volumes (i.e. corrected for porosity). In addition, since sand and gravel have different transport characteristics and ultimately different fates when they enter the coastal system, it is important to treat them separately for sediment budgeting. To determine the relative proportions of sand and gravel involved in river to coast sediment

transfers, bulk sampling of the surface and sub-surface river sediment was undertaken following the criteria set out in Section 4.4. This section presents detailed results and analysis of all surface and sub-surface sediment data (see Figure 4.1 for location of all sediment samples).

5.3.1 Surface characteristics

The distal 6km reach of the lower River Spey has a mean surface D_{50} of 42mm (range 22-60mm) and displays no systematic downstream change in sediment size (Figure 5.12). The results of all surface samples are presented in Appendix C. The samples of 400 surface clasts at bar-head locations have a lower mean D_{50} (34mm or -5.08ϕ) compared to that based on random 100 clast samples (mean D_{50} is 42mm). Sorting (σ_g) of all samples lie within the range 1.7-1.9 indicating relatively well-sorted sediment.

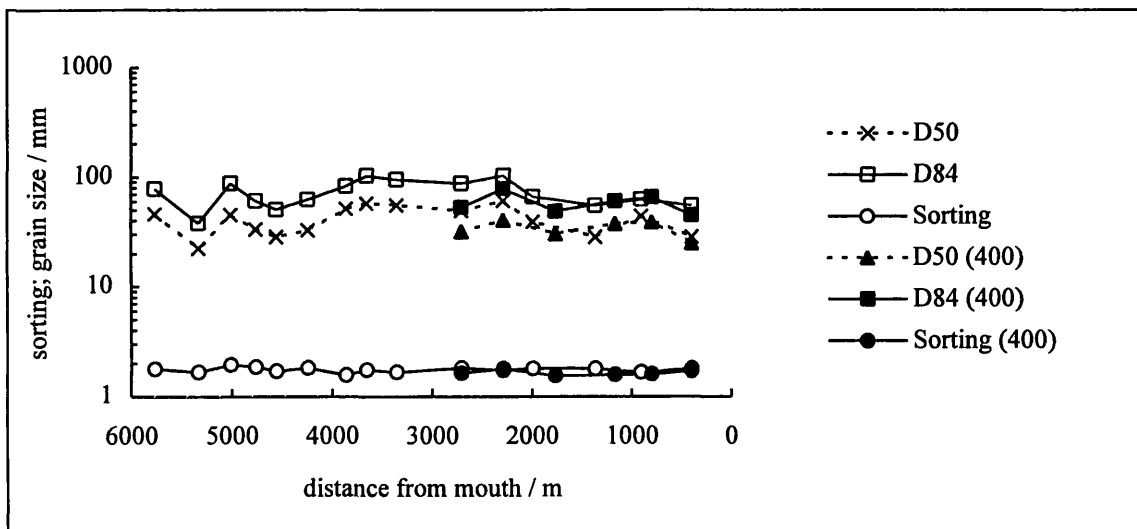


Figure 5.12: Surface grain-size characteristics of the lower River Spey gravels (> 8mm). The graph shows the lower values of D_{50} and D_{84} recorded using a surface sample of 400 clasts in a small 0.5 m^2 area of the bar-head, compared to those recorded based on a random (pacing) sample of 100 clasts.

The surface sand is a coarse to medium sand, with the D_{50} ranging from 0.3 to 0.7mm (Figure 5.13) and a mean D_{50} of 0.5mm. No systematic downstream fining of the surface sand exposures was identified (Figure 5.13), although grain sizes vary in response to local conditions. The sands are relatively well-sorted ($\sigma_g = 1.4$ -1.7) with the exception of the sample collected ca. 1400m upstream ($\sigma_g = 2.78$) which contained a mix of sand and more granular sediment.

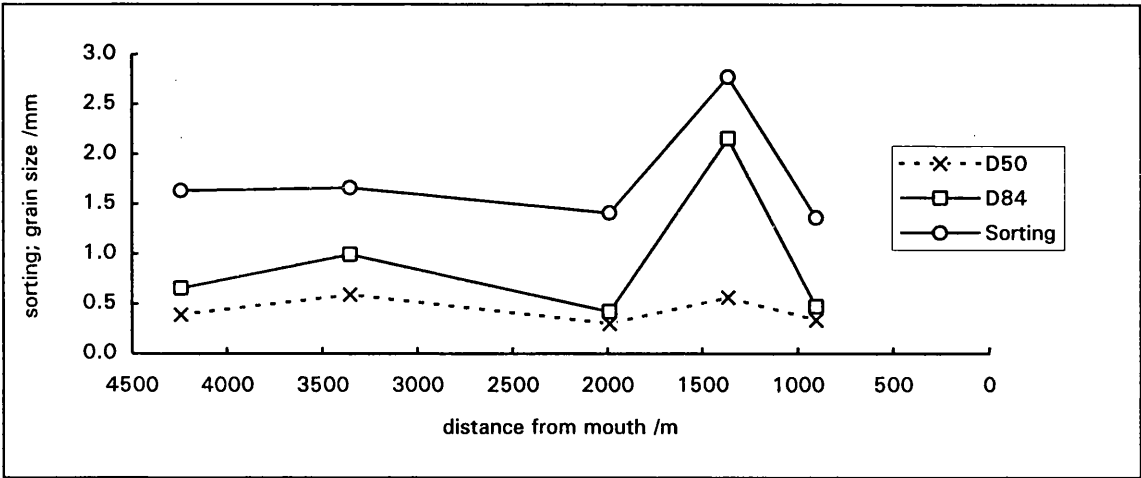


Figure 5.13: Surface grain-size characteristics of the lower River Spey (<2mm)

5.3.2 Bulk samples

The bulk sample sediment results are presented in Appendix D. Both surface and sub-surface grain size distributions at all sites are bimodal (Figure 5.14). The D_{50} of surface sediment ranges from 18 to 46mm, with a mean of 31mm; the sub-surface sediment has a lower D_{50} (15 to 35mm) in all cases (Figure 5.15).The percentage of sand ($d < 2\text{mm}$) in the surface samples varies from 7 to 23% and shows no systematic downstream trend (Figure 5.16). Sub-surface samples contain a greater proportion of sand (12 to 28%).

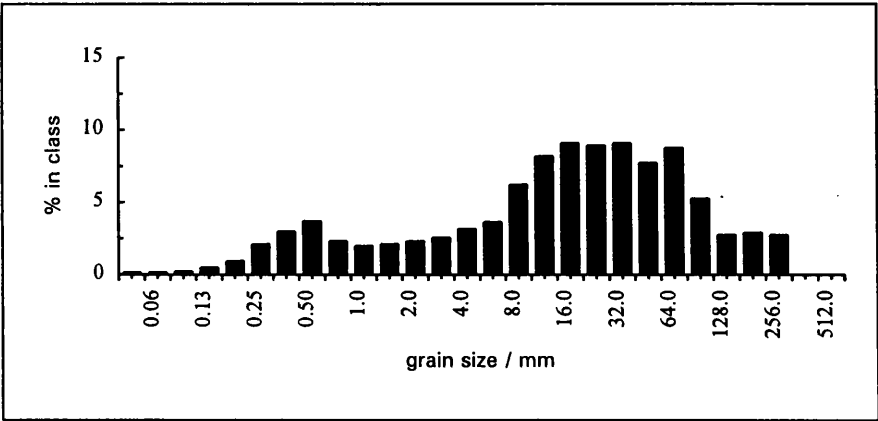


Figure 5.14: Surface grain size distribution at Essil (total sample mass is 1017kg). $D_{50} = 14\text{mm}$, $D_{84} = 82\text{mm}$ and sorting $= \sqrt{(D_{84} / D_{16})} = 6.94$

Church et al. (1987) note that when sediment $< 2\text{mm}$ comprises less than 30% of the bed material it occurs as interstitial fill. This condition is met in all surface and sub-surface samples, indicating that the sand is likely to be transported in suspension in this reach, settling out as interstitial fill at low flows. Consequently, in large floods substantially larger volumes of such material may move through the reach and enter the coastal system than are trapped here.

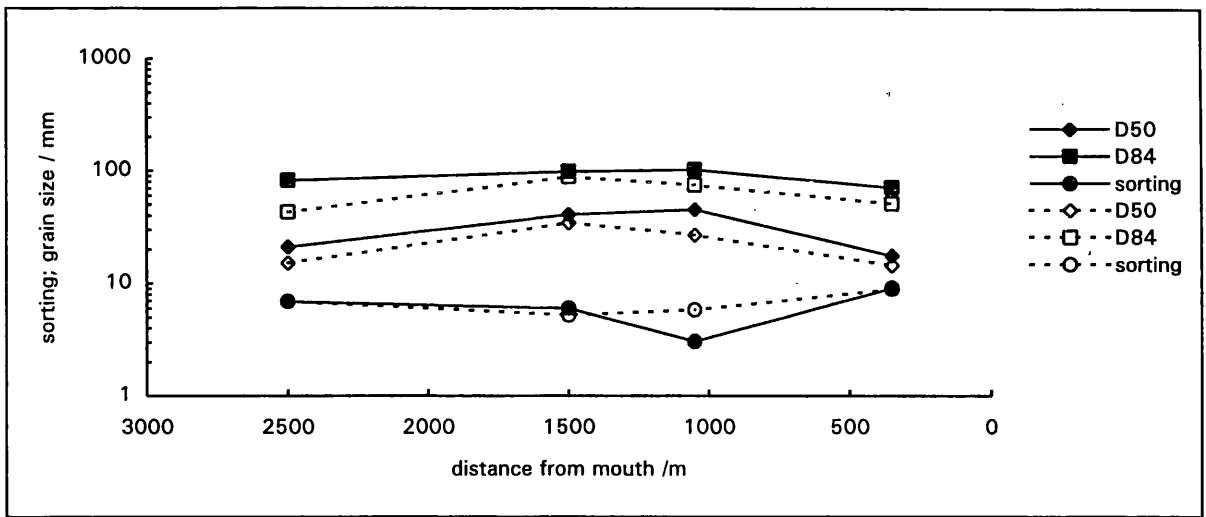


Figure 5.15: Surface (in bold) and sub-surface grain size characteristics of the lower River Spey (bulk samples).

The bulk samples at 2500m and 350m upstream contain the highest proportion of sand and have the lowest D_{50} (Figures 5.15 and 5.16). The upstream sample was collected at the flood infill deposit at Essil. The poor sorting, large range of grain sizes and high proportion of sand in this sample (Figure 5.15) reflects the depositional history of this sediment. The downstream sample was collected from a tidally influenced bar, close to the mouth, which may account for the finer grain-sizes and high proportion of sand.

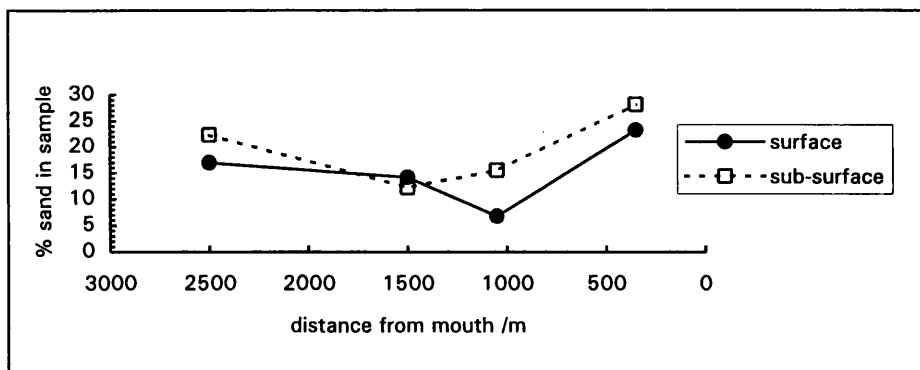


Figure 5.16: Percentage of sand in surface (in bold) and sub-surface bulk samples.

The size distribution of the mean annual sediment yield of gravel-bed rivers tends to be similar to the sub-surface rather than the surface sediment (Parker 1990; Martin and Church 1995). Therefore, volumes of fluvial sediment entering the coastal system are converted into sand and gravel fractions using the distributions acquired from the 4 bulk samples of sub-surface sediment. As there is no downstream trend in the percentage of sand in each sample, the mean percentage of sand ($20 \pm 3\%$) is used. It is assumed that the volumetric transfers quantified in the lower Spey (Section 5.4.2) comprise 20% sand and 80% gravel.

5.3.3 Sediment porosity

Since bedload transport rates are generally given as mineral volumes (or weights) per unit time, each volume change estimate is multiplied by $(1-\rho)$ where ρ is the porosity of the material (equation 2.3, Section 2.1.1.2). Porosity is a difficult measurement to make in the field (Martin and Church 1995) and is estimated using published data. Komura (1961) and Carling and Reader (1982) show that porosity for alluvial gravels within the size range of those of the Spey fall consistently in the range 0.2 - 0.3. The poor sorting and hence high packing of sediments in the upland Pennine streams sampled by Carling and Reader (1982) give lower values of porosity than those reported by Komura (1961) for larger rivers. Lane et al. (1995) used a value of 0.2 for the poorly sorted bed-material of an Alpine pro-glacial stream, calculated using the empirical relationship between median particle diameter and porosity provided by Carling and Reader (1982). The Spey sediments are well sorted, so this empirical relationship may not be valid and the upper value of 0.3 ± 0.05 for porosity is used (Komura 1961). This value is consistent with that used by Wathen (1995) for the Allt Dubhaig bed material, which has similar D_{50} and sorting characteristics to the Spey. Choosing the correct value for sediment porosity is important and can potentially introduce large errors in the sediment transport estimates (see Section 5.4.4).

5.4 Short-term (contemporary) sediment budget

5.4.1 Empirical estimates of bedload transport

While it is acknowledged that estimates of bedload transport from empirical formulae may be problematic (see the discussion in Chapter 2) they are used here to (a) provide a check on the validity of the sediment transfers quantified herein, and (b) provide an upstream boundary condition (Q_b) for input into the sediment budget equation (Section 2.1.1.2, equation 2.3). Two estimates are available:

(1) Empirical estimation using local hydraulic data at a section, Parker's (1990) bedload transport relation and flow data

This method can be used to derive a general bedload rating curve for the River Spey of the form:

$$Q_b = a(Q - Q_c)^b \quad (5.4)$$

where Q_b is the bedload transport rate ($\text{m}^3 \text{s}^{-1}$), Q and Q_c are the discharge and threshold discharge for initiation of transport ($\text{m}^3 \text{s}^{-1}$), respectively, and a and b are empirical constants ($b > 1.5$). The surface-based bedload transport relation of Parker (1990) was used

to estimate bedload transport at a range of discharges. The problems of using empirical bedload relations are well known (e.g. Davies 1987; Gomez and Church 1989). However, when used carefully, and when calibrated and tested, the estimates are of value and provide a low-cost technique to estimate bedload transport. The bedload rating curve derived here was combined with continuous flow data (recorded every 15 minutes) from Boat o'Brig (ca. 10km upstream of the study reach) to estimate the annual sediment supplied to the mouth in each year of the study.

A stage-discharge (rating) curve was derived from the geometry of a stable cross-section (section 14, Figure 5.6), the local energy slope (S) and the D_{84} of surface sediment as follows. Surface roughness was calculated using the composite equation of Bray and Davar (1987):

$$\frac{1}{\sqrt{f_f}} = a \log(R / D_{84}) + b \quad (5.5)$$

where f_f is the Darcy-Weisbach friction factor, R is the hydraulic radius (m) and a and b are empirical constants, estimated as 2.0 and 1.1, respectively (Bray and Davar 1987). The rating curve was calculated using equation (5.5) in conjunction with the definition equation:

$$\bar{u} = (8gRS / f_f)^{0.5} \quad (5.6)$$

and the continuity equation. In 5.6, \bar{u} is the mean flow velocity (ms^{-1}) and g is the acceleration due to gravity (9.81ms^{-2}). Shear velocities (u^* , ms^{-1}) were calculated using:

$$u^* = \sqrt{gRS} \quad (5.7)$$

Parker's (1990) ACRONYM 1 programme was used to estimate volumetric transport rates per unit width at different shear velocities using the bedload transport equation developed in Parker (1990). The equation is based on a reanalysis of data from Oak Creek, Oregon, to produce a relationship based on surface grain sizes. The bulk grain size distribution of surface sediment ($> 2\text{mm}$) at section 14, the submerged specific gravity (for quartz = 1.65) and shear velocities corresponding to twenty discharge increments from the rating curve were input into ACRONYM 1. The unit width volumetric transport rates (in m^2s^{-1}) were converted to volumetric transport rates (m^3s^{-1}) by multiplying by the channel width at each discharge.

The bedload rating curve (Figure 5.17) for the Spey is fitted by equation 5.4 with the constants a , b and Q_c having values of 2.22×10^{-11} , 3.13 and $0.42 \text{ m}^3\text{s}^{-1}$, respectively.

$Q_c=0.42$ is a statistical best fit and is not a real Q_c for bedload initiation. The curve is statistically significant ($p < 0.05$, adjusted $R^2 = 0.998$).

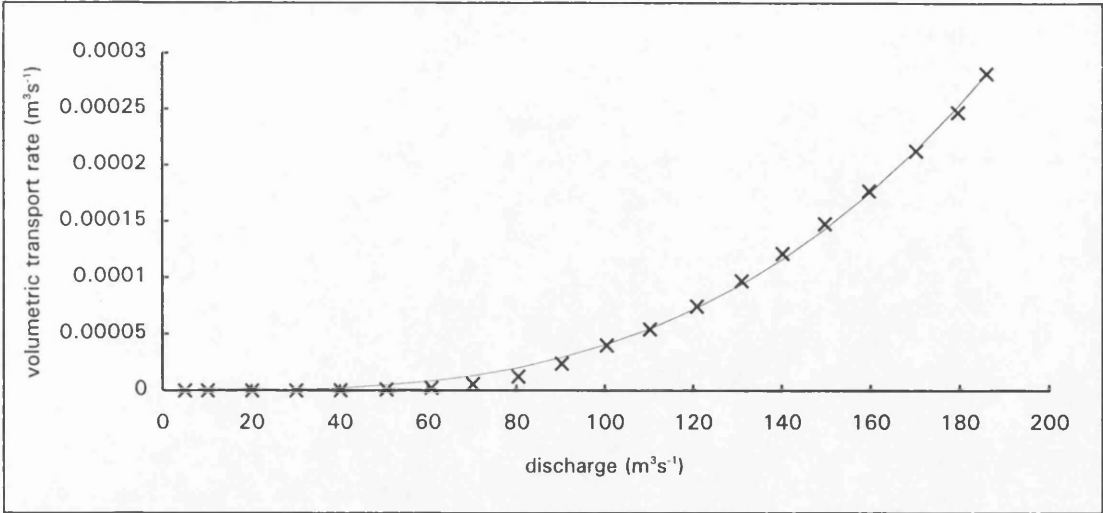


Figure 5.17: Bedload rating curve for the lower River Spey (at section 14). The data is described by the equation $Q_b = a(Q - Q_c)^b$ where a and b are 2.22×10^{-11} and 3.13, respectively (see text for details).

The estimated bedload (Table 5.7) closely follows the peak flows, with high values of bedload transport estimated during the floods of September 1995 and July 1997 (Section 3.3.1). The bedload volume estimated in each quarter varies from 1978 m³ (Jan - Mar, 1993) to 6 m³ (July -Oct, 1996) highlighting the sensitivity of bedload to flow conditions. The high amounts of bedload transport in the first quarter of 1993 and 1994 correspond to the high flow conditions at that time (Section 3.3.1).

	1993	1994	1995	1996	1997	1998	1999
Jan-Mar	1978	1345	505	170	855	245	514
Apr - June	143	477	159	47	102	310	261
July - Oct	18	12	1455	6	1031	178	55
Oct -Dec	781	240	179	522	160	699	72*
Annual load	2920	2074	2298	745	2149	1432	901*

Table 5.7: Bedload transport (in m³) estimated using the empirically derived bedload rating curve $Q_b = 2.22 \times 10^{-11}(Q - 0.42)^{3.13}$ and continuous flow data. * the 1999 annual load is incomplete as flow data was only available up to 8/11/99.

The derived bedload rating equation was combined with the flow duration curve for 1953-1998 (Figure 3.11), in order to estimate the average annual sediment load to the coast over the entire period of flow record. The 1953-1998 curve was constructed using daily mean values (DMV) and not continuous flow data, so a correction factor was applied. The correction factor is derived as the ratio between total bedload in 1995-1998 estimated firstly using continuous data (6653 m³, Table 5.7) and secondly using the 1995-1998 flow duration curve (based on DMVs) (4614 m³). It is assumed that this factor (1.43) can be

used to convert bedload estimates derived from DMV data to that from continuous flow data. This correction, although rather crude, gives an average annual sediment load to the coast of 1834 m^3 over the period 1953 to 1998. This does not differ significantly from the annual loads estimated over the study period (Table 5.7).

One of the problems of using this rating curve to estimate bedload transport is that it assumes bedload is controlled entirely by flow hydraulics, and fails to consider other important variables such as sediment supply (Richards 1982). In addition, it is recognised that rating curves underestimate total loads (Ferguson 1986, 1987). An unbiased estimator of the true load is given as (from Ferguson 1986):

$$L_{cr} = L_r e^{2.651s^2} \quad (5.8)$$

where L_{cr} is the corrected load, L_r is the rating curve load and s is the SE of the rating curve estimate. In this case, the rating curve is calculated using continuous discharge data from which bedload is predicted, so $s \approx 0$ and $L_{cr} \approx L_r$. However, in reality s is likely to be greater than 0, adding a further source of error from empirical curves. Given that $b = 3.13$ in equation 5.4, s may be of the order of ≈ 0.5 , so substituting in equation 5.8, $L_{cr} = L_r \times 1.94$. This suggests that the empirically derived estimates of bedload transport (Table 5.7) may be underestimated by a factor of around 2. Nevertheless, these estimates give values for the boundary condition of sediment input (Q_i) (equation 2.3, Section 2.1.1.2) and can be compared to estimates obtained using the morphological approach (Section 5.4.2).

(2) Estimates obtained using MIKE 11 (Dobbie & Partners 1990; Riddell and Fuller 1995)

Detailed river modelling studies of the lower River Spey were carried out by Dobbie & Partners (1990) using the MIKE 11 numerical model, developed by the Danish Hydraulic Institute (1990). The model was run for a series of flood events of different return periods to predict the volume of sediment deposited at Speymouth during entire flood events. The results show that a flood with a 1 year return period may deposit between 7 300 and 10 250 m^3 of sediment at the mouth (Figure 5.18). The average annual quantity of sediment transported by the river was estimated to be around 8 000 m^3 (Dobbie & Partners 1990), although this is variable depending on the flow conditions in any given year. The MIKE 11 model was run for the period July 1990 to May 1992 and predicted a supply of $22\,500 \pm 2\,500 \text{ m}^3$ of sediment from the river to the coast over this period (Riddell and Fuller 1995).

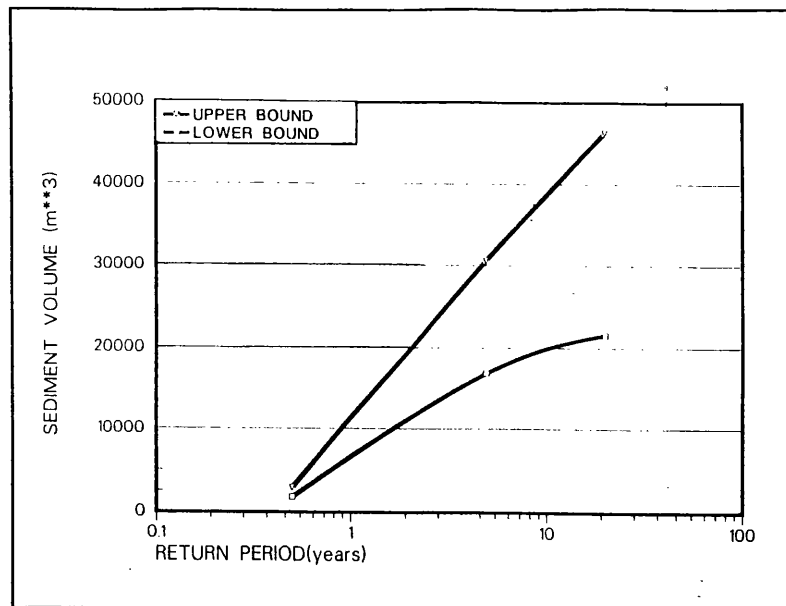


Figure 5.18: Volume of sediment (in m³) deposited at Speymouth during flood events, calculated by river modelling studies carried out by Dobbie & Partners (1990). The lower estimate indicates the volume of material which is transported along the channel, while the upper estimate includes the volume of material which is eroded locally due to constriction of flow at the mouth.

5.4.2 Quantification of the fluvial sediment budget for the Lower River Spey

The sediment transport rate is computed using equation 2.3, Section 2.1.1.2:

$$Q_o = Q_i - (1 - \rho) \delta S / \delta t \quad (2.3, \text{chapter } 2)$$

Net storage changes for each reach ($\delta S / \delta t$) are quantified using planimetric methods and presented in Section 5.2.3 (Figure 5.11) and porosity (ρ) is 0.3 ± 0.05 (Section 5.3.3). To construct the sediment budget using this method requires measurement or assumption of a transport rate, Q_i (Section 2.1.1.2). As direct measurement of gravel transport in a large river, such as the Spey, is problematic, Q_i is estimated using the bedload rating equation derived in Section 5.4.1. These estimates of Q_i include the gravel component of bedload only and are in mineral volumes (i.e. corrected for porosity). All estimates are increased to include the sand component of sediment transfers (assuming that material is 80% gravel and 20% sand, Section 5.3.2). This assumption and the inconsistency of predicting transport rates using empirical formulae (Gomez and Church 1989) may introduce significant errors into the transport rates derived herein, especially as errors propagate downstream.

Sediment budgets for each period (Figure 5.19) are given in Appendix B. Transport rates vary significantly between periods and possible explanations for this are explored in Section 5.4.3. Transport rates are highest in 1993-94 when $41\,260 \text{ m}^3 \text{ a}^{-1}$ of sediment was transported to the coast (i.e. Q_o of downstream reach 11). The high transport rates recorded

during this period reflect both the high flow conditions, but more importantly, the channel avulsion which occurred in January 1994 (Section 5.2.3). This avulsion released large volumes of previously stable deposits (e.g. vegetated floodplain) for potential transport and greatly affected all reaches downstream of reach 6 (Figure 5.7), which show a 3-fold increase in transport rates (Figure 5.19).

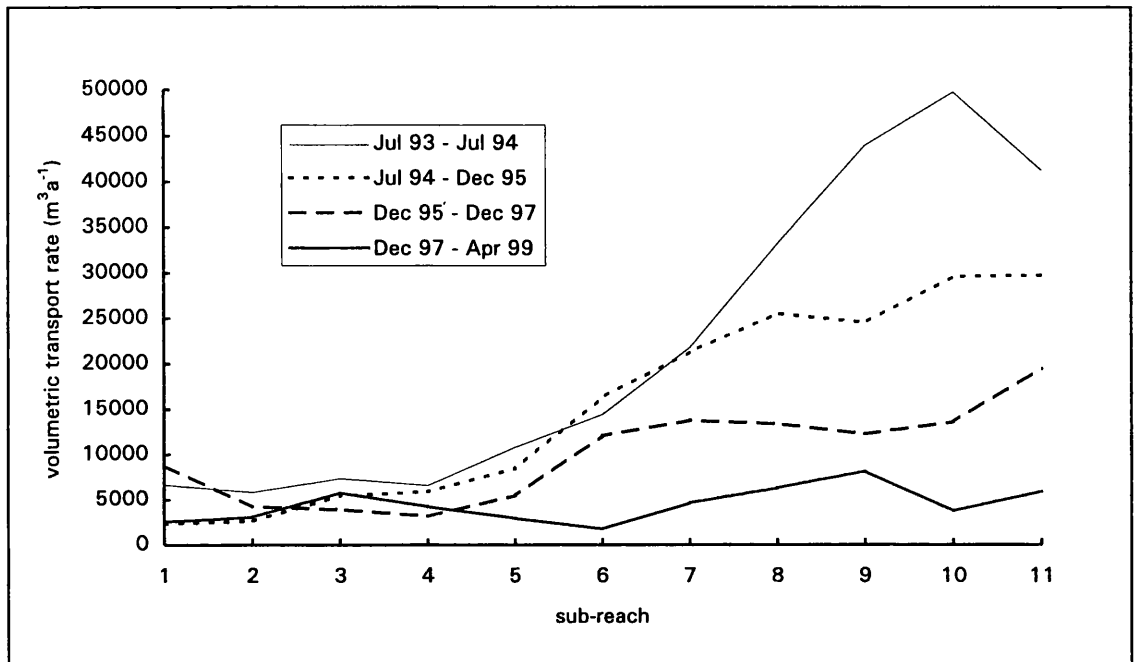


Figure 5.19: The sediment budget along the lower River Spey represented as sediment transport at sub-reach boundaries. The downstream boundary of sub-reach 11 is the Speymouth delta.

Transport rates remained high between July 1994 and December 1995 with ca. $30\,000\text{ m}^3\text{ a}^{-1}$ of sediment reaching the coast. The combination of inherent channel instability, following the channel change early in 1994, and the flood of September 1995 may account for this. Again, transport rates increase significantly downstream of reach 6 reflecting the higher activity and instability of the lower reaches of the Spey.

Transport rates are reduced in the period December 1995 to December 1997, with ca. $20\,000\text{ m}^3\text{ a}^{-1}$ of sediment entering the coastal system (Figure 5.19). Although the flood of July 1997 caused morphological change in the reach, channel changes were less extensive than past floods and had less effect on the transport rate, perhaps due to increasing vegetation cover and stability of the bar deposits.

The former Essil channel was infilled during this period (Section 5.2.3) providing a direct estimate of Q_i , which is calculated as the volume of infill less the net volumetric changes in the two upstream reaches (Figure 5.9). Errors in this estimate of Q_i are quantified in Section 5.4.4. Transport rates into reach 1 (Q_i) are high (ca. $7\,000\text{ m}^3\text{ a}^{-1}$), but fall

dramatically as a result of net deposition in reaches 2, 3 and 4 (Figure 5.19). The loss of the large volume of sediment (ca. 25 000m³ bulk) from the system to Essil channel infill may be an additional contributory factor to the increased stability in the downstream reach.

Lack of significant morphological change between December 1997 and April 1999 (Figure 5.10) led to the lowest transport rates recorded in this study, with an estimated 6 000 m³ a⁻¹ of sediment reaching the coast during this period. The rate varies down the river reaching a minimum of 1 750 m³ a⁻¹ in reach 6 (Figure 5.19).

5.4.3 Environmental factors

This section examines the relationship between the estimated transport rates and river flows during the study period. Estimates of the initial Q_i for each period are directly related to continuous discharge by the bedload rating equation 5.4. Transport rates quantified using the morphological approach are more likely to be related to the occurrence of flows capable of initiating morphologic change.

A critical threshold flow, above which morphologic change is likely to occur needs to be identified for the lower Spey. In other rivers where there is an absence of data relating flows to bedload transport, the mean annual flood ($Q_{2.33}$) or the median flood ($Q_{1.5}$) is often used. For the lower Spey, these are 695 m³ s⁻¹ and 485 m³ s⁻¹, respectively (Section 3.3.1, Figure 3.10). Analysis of flows over the study period (Table 5.8) indicate that $Q_{1.5}$ was exceeded on only 5 occasions and was not exceeded in the period July 1993- July 1994, when the highest sediment transport rates (and most morphological changes) were recorded. This indicates that the threshold flow on the lower Spey is less than $Q_{1.5}$ or that the channel is sensitive (and thus susceptible to change) for some other unknown reason (e.g. sediment accumulation, accessibility to sediment, channel orientation).

	Jul 90-May 92	Jul 93-Jul 94	Jul 94-Dec 95	Dec 95-Dec 97	Dec 97-Apr 99
transport rate (m ³ a ⁻¹)	12 273	41 261	29 773	19 553	5 932
days mean discharge > 485 m ³ s ⁻¹	1	0	3	2	0
days mean discharge > 280 m ³ s ⁻¹	5	11	7	9	3
peak flow (m ³ s ⁻¹)	534	460	718	705	363

Table 5.8: Annual sediment transport rates to the coast compared to flow conditions. The 1990-1992 transport rate is from numerical modelling (Riddell and Fuller 1995).

The dominant or effective discharge can be defined as the flow which does the most work, where work is defined as sediment transport rate (Wolman and Miller 1960). The effective discharge for the lower Spey was obtained using the flow duration curve from 1953 - 1998 (Figure 3.11) combined with the bedload rating curve (Figure 5.17). From this analysis, the

effective discharge lies in the range $240\text{--}290\text{ m}^3\text{s}^{-1}$ (Figure 5.20), which is roughly equivalent to the 0.5 year-flood on the lower Spey (Section 3.3.1).

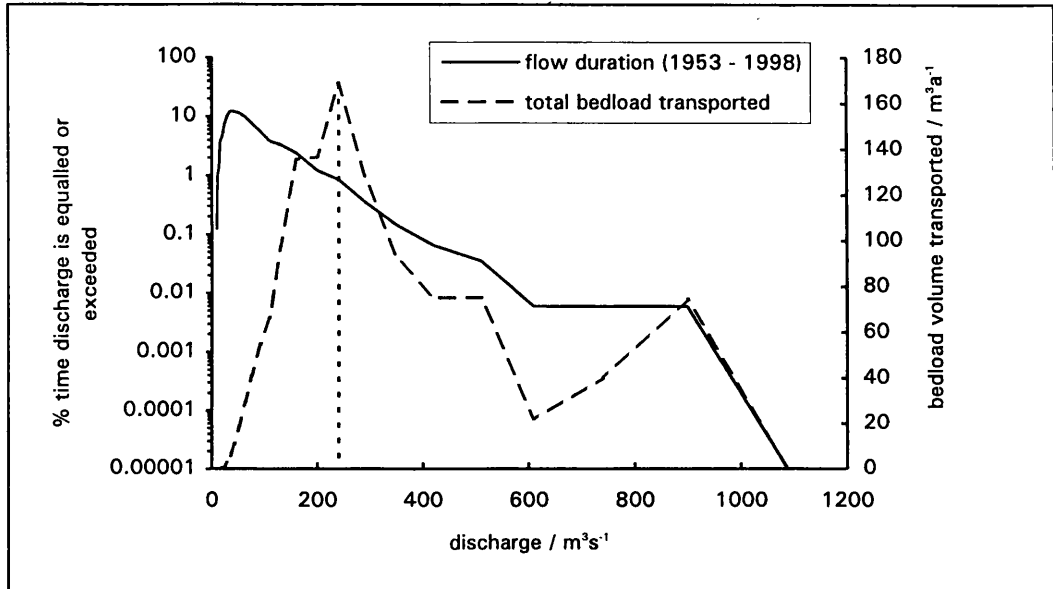


Figure 5.20: 1953-1998 flow duration curve combined with the bedload rating curve to estimate the effective discharge (i.e. the discharge at which most sediment transport occurs) on the lower River Spey. The effective discharge for the transport of bedload lies in the range $240\text{--}290\text{ m}^3\text{s}^{-1}$.

An alternative determination of the threshold flow for morphological change on the Spey is to analyse flow conditions during times of past morphological change. The channel avulsion of January 1994 occurred during a flood with a peak discharge of $280\text{ m}^3\text{s}^{-1}$ (Stratton, 1996, pers comm.). This flow agrees closely with the estimate of the effective discharge and is adopted herein as the threshold flow for initiation of morphological change. It is acknowledged that this flow may not cause change every time it is exceeded, as changes may be related to other factors inherent in the river system.

There is a strong relationship ($r = 0.94$, $p < 0.05$) between the number of days per year when the threshold flow is exceeded and the annual transport rate (Figure 5.21). The highest annual transport rate (1993-1994) occurred when flows of $280\text{ m}^3\text{s}^{-1}$ occurred on 11 days per year and the lowest transport rate (1997-1999) corresponds with a frequency of occurrence of only 2.1 days per year. The 1990-1992 transport rate for the lower Spey is derived from numerical modelling using the Mike 11 model (Section 5.4.1, published in Riddell and Fuller 1995) and correlates well with the rates obtained herein (Figure 5.21). The transport rates obtained using the bedload rating curve (Section 5.4.1) predict much lower values of bedload transport than the morphological approach and show little variation with the frequency of events $> 280\text{ m}^3\text{s}^{-1}$, although they are strongly correlated ($r = 0.94$).

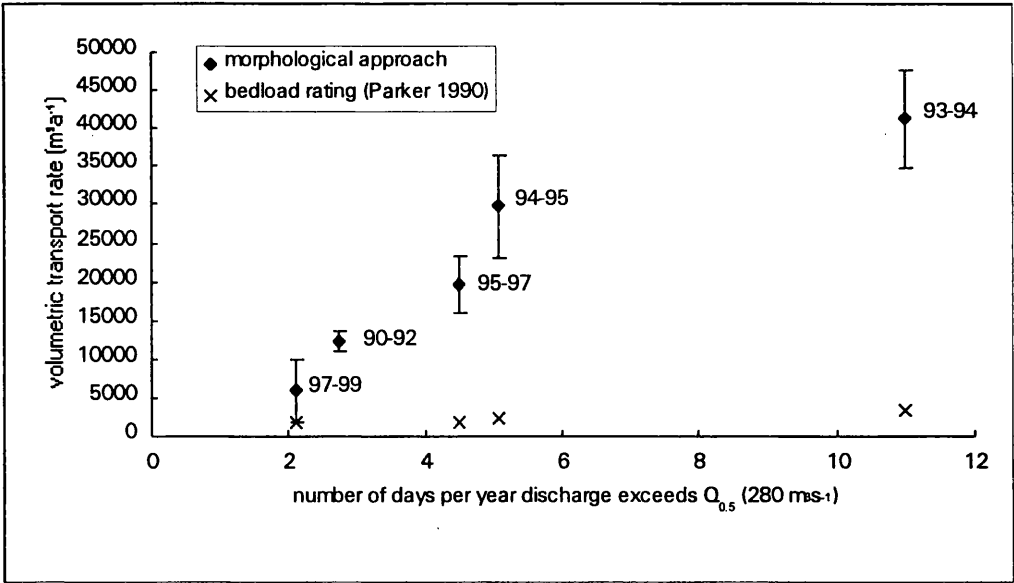


Figure 5.21: Relationship between the volumetric transport rate out of the downstream reach and the number of days per year the threshold flow ($280 \text{ m}^3 \text{ s}^{-1}$) is exceeded. Transport rates obtained from the bedload rating curve (section 14) are shown for comparison.

Empirical estimates of bedload transport are closely related to continuous flow and fail to account for other important variables, such as sediment supply or major system adjustment. The bedload rating equation is derived from flow conditions and local hydraulics at a stable section, and therefore fails to account for major morphological shifts and associated rates of sediment transport which can occur in wandering river channels.

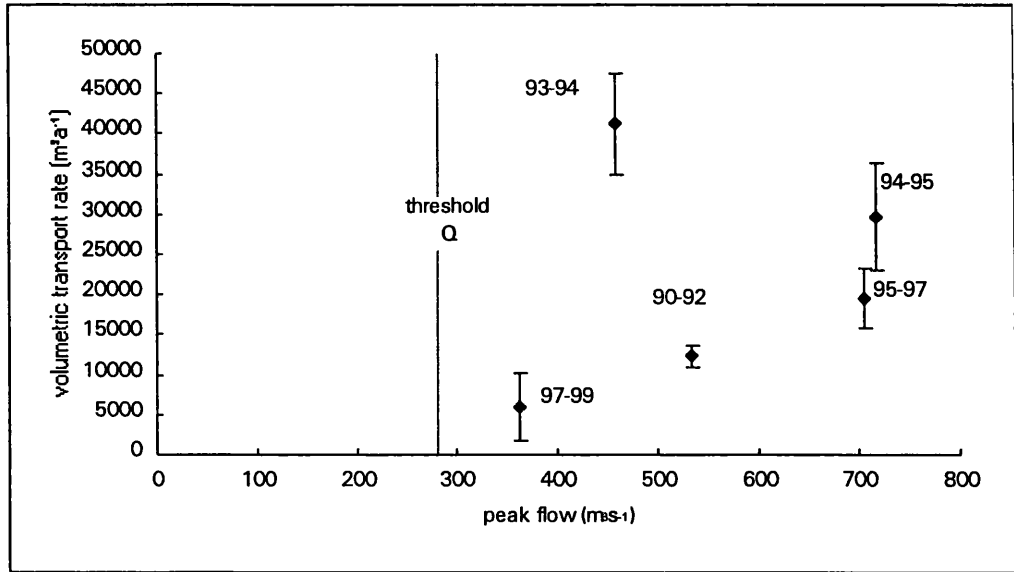


Figure 5.22: Annual transport rates at the downstream reach compared to the peak flow recorded during each period.

The estimated annual sediment transport to the coast is compared to the peak flow recorded in each period (Figure 5.22). The data shows considerable scatter, although if the 93-94 rate is omitted there is a positive relationship between peak flow and annual transport rate.

The highest annual transport rate was recorded in 1993-1994 during which the peak flow was relatively low ($460 \text{ m}^3 \text{ s}^{-1}$), although there were a large number of events exceeding $280 \text{ m}^3 \text{ s}^{-1}$. Channel instability caused by the avulsion in January 1994 also contributed to the high rate.

This analysis indicates that transport rates are not directly related to flow conditions, as assumed by simple bedload rating curves (Richards 1982). Other factors such as sediment supply, sediment accessibility and channel stability are important controls of bedload transport rates in wandering gravel-bed rivers. Channel instability caused either by natural avulsion or dredging (cf. Martin and Church 1995) provides access to previously stable sediment, increasing transport rates during subsequent floods. Therefore it is not always the magnitude of the flood that influences rates of sediment transport, but the stability of the channel and bars prior to the flood. For example, the large areas of bare gravel following the avulsion in January 1994, led to high transport rates in the succeeding periods, as the channel was able to freely respond to high flow events (e.g. 1994-1995). As the channel begins to stabilise, vegetation establishes on bars and islands and the system becomes less sensitive to flow variation (e.g. 1997-1999). Less morphological change is likely as the channel stabilises, resulting in lower transport rates.

5.4.4 Error analysis

The accuracy of sediment transport estimates is difficult to assess since the true sediment transport rate is unknown. Availability of comparative data remains a problem in all studies which use the morphological approach (Ashmore and Church 1999) and is one of the reasons why error analysis is required.

Errors in the sediment porosity assumption, the estimation of Q_i and the reach storage change estimates, are the principal sources of uncertainty in the budget estimates. Following Martin and Church (1995), the porosity estimate was assigned an error range of ± 0.05 . The uncertainty in Q_i is potentially quite large, given the generally poor performance of empirical bedload relations (Gomez and Church 1989) and the tendency for rating curves to underestimate total loads (Ferguson 1986). It is assumed herein that Q_i has a maximum uncertainty of $\pm 100\%$. For December 95- December 97, the error in Q_i is less and equal to the uncertainty in the quantification of the volume of sediment infill at Essil.

The uncertainty in the estimation of reach storage change is a combination of errors, which occur at several stages of analysis. These include mapping errors, digitising and overlaying errors, errors in the estimation of d (the depth of mobile sediment) and errors in the

application of the water level correction (δw). The RMS error in each polygon of change is computed in the GIS and combined to quantify the absolute errors (in m^3) of storage change estimates for each reach (Appendix E).

Mapping errors

As a compromise between accuracy and the area covered in a given time, field maps were compiled by pacing out the bar and channel changes using either the 1994 1:5 000 photography or the 1995 1:10 000 geomorphology map as a base (Section 4.1.3). This minimises potential errors when mapping change. Any changes were paced out in the field from fixed markers and known points, with the aid of numerous aerial photographs and maps to increase accuracy. To estimate potential mapping errors, consider a bar change of 50m over a length of ca. 300m. Typically, this change would be paced out in the field 5 times at different locations on the bar. Assuming each repeated pacing quantifies the distance to within $\pm 5\text{m}$, the mean RMS planimetric error is 2.23m. Where aerial photographs are used to map the channel (July 1993 and 1994) a planimetric error of 2m is assumed.

Digitising and overlay errors

Digitising and overlay errors were estimated by measuring the mean planimetric error in the co-ordinates of Ordnance Survey control points in the map overlays. Fourteen OS control points were digitised and used to overlay successive maps (Section 4.1.3). Mean x and y displacements between overlays give planimetric errors of 0.36m, 0.52m, 0.37m and 0.69m for the periods 1993-1994, 1994-1995, 1995-1997 and 1997-1999, respectively. Planimetric errors from digitising and overlaying maps were minimised during analysis (Section 5.2.3); for example, if the map overlay recorded change in a particular bank or bar where it was known that no actual change had occurred this change was omitted from the storage change estimates.

Mapping, digitising and overlaying errors are combined to give the RMS planimetric error for each period. Planimetric errors can be compensating and are sometimes discounted completely from error estimation (e.g. Ham 1996) as it is assumed that a shift in channel position caused by placement error will result in apparent erosion of one bank or bar and deposition on the other. This is an oversimplification and planimetric errors are included here. Planimetric errors (δ) in m are converted to area errors (in m^2) as they are related to the area of each bar or bank change (A) by:

$$\text{error } (m^2) = (\sqrt{A} + \delta)^2 - A = 2\delta\sqrt{A} + \delta^2 \quad (5.9)$$

The error term (m^2) for each polygon of change is converted into a volume error (m^3) by multiplying it by the appropriate depth of mobile sediment (d) for the particular bar or bank (Section 5.2.3).

Storage Volume errors (uncertainty in the depth of mobile sediment)

Volume-area ratios (d) were computed for all surveyed bars in Section 5.2.2. The mean survey point spacing is 7.62m (excluding the densely surveyed bars used for spacing error analysis, Section 4.6.1). A mean error of ca. 1% in bar storage volume is associated with this spacing (Figure 4.5, Section 4.6.1.5). An additional error of 0.5% (ca. 1 SE) is assumed to apply where the mean depth of mobile sediment for bars or banks is used. For each volume of change the absolute volumetric error associated with the depth of mobile sediment is calculated as 1.5% of the volume change.

Uncertainty in the water level correction

Water level corrections were applied to reaches downstream of reach 6 in the periods 1994-95 and 1995-97, due to discharge variation between surveys (Section 5.2.3). It is difficult to assess the error in the estimation in δw (equation 5.2) and it is assumed herein that the estimation is correct to within $\pm 20\%$. Errors are computed using this range of width estimates to compute the absolute error (in m^3) associated with the water level correction (Appendix E).

Errors propagate down the river system, as the transport rate out of each reach (Q_o) is used as the transport rate into the next reach, Q_i (equation 2.3). Errors in the transport rate for reach n are calculated using (following Martin and Church 1995):

$$E_n = \sqrt{(\delta Q_i)^2 + (\delta \rho_n)^2 + (\delta \Delta S_n)^2} \quad (5.10)$$

where δQ_i is the uncertainty in the transport rate into reach n , $\delta \rho$ is the uncertainty of the porosity and $\delta \Delta S$ is the uncertainty of the storage change estimate for reach n .

Cumulative errors in Q_o are given in Appendix E and are presented graphically in Figure 5.23. Most of the uncertainty in transport rates are due to errors in Q_i (Figure 5.23), which increase downstream because of error propagation in the cumulative calculations. The error in the transport rate at reach 11 (i.e. as the sediment reaches the coast) is up to three times greater than the error range of the initial Q_i . Absolute errors are greater for the periods

1994-95 and 1995-97, as these include the error in the application of the water level correction.

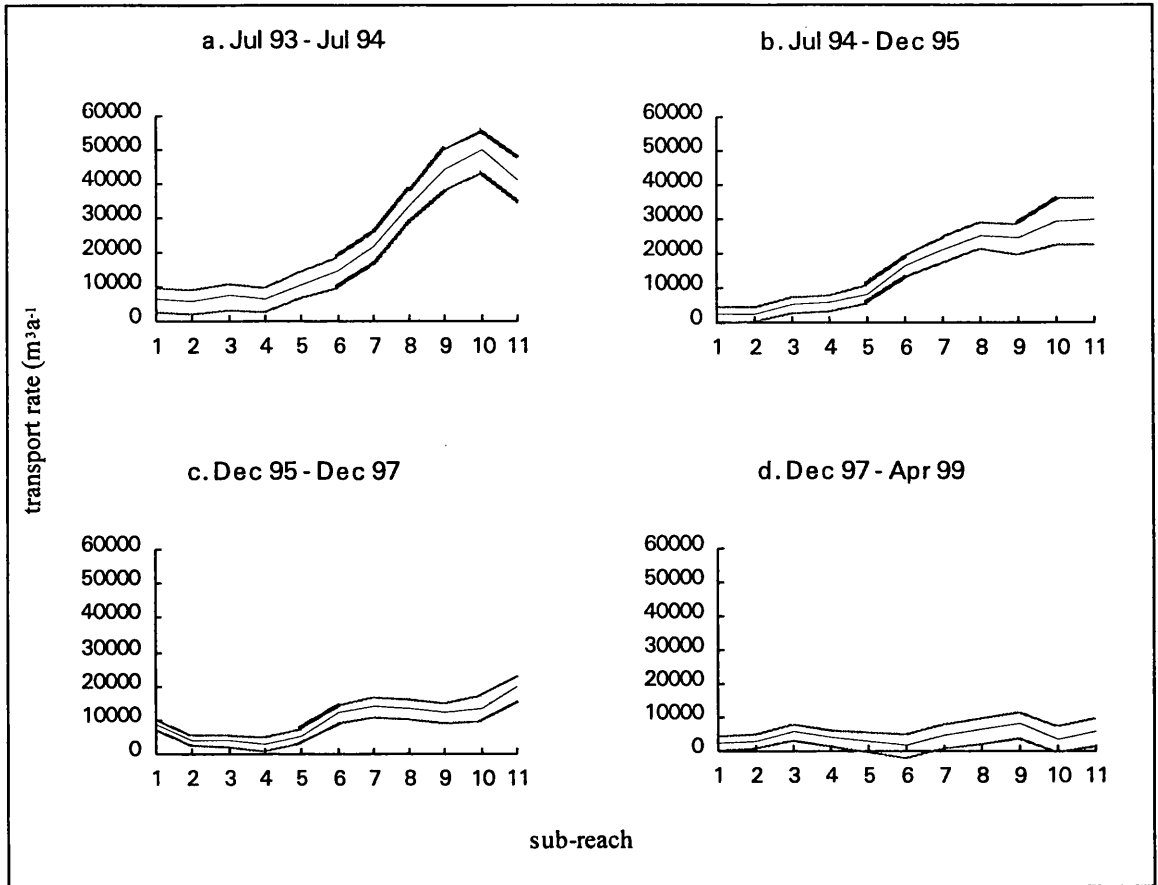


Figure 5.23: Sediment transport rates for the lower River Spey with estimated error ranges for the periods: a. July 1993 - July 1994; b. July 1994 - December 1995; c. December 1995 - December 1997; and d. December 1997 - April 1999, calculated from planimetric changes.

The errors are considered acceptable given the magnitudes of the transport rates entering the coastal system, being of the order 15-22% of the estimated transport rate in all periods except the low transport period 1997-1999 when it may have been 69% (Figure 5.23d). This method can be considered reliable for periods with high transport rates (and large amounts of morphological change), although in periods with only minor morphological change (e.g. 1997 - 1999) the errors involved in quantifying sediment transfers may be larger than the transfers that are actually occurring. In such cases, the technique gives order-of-magnitude estimates of sediment entering the coastal system.

5.5 Delta

The fluvial sediment transfers quantified above reach the coast via the Speymouth delta and are redistributed in the coastal zone by the interaction of fluvial and coastal processes.

This section investigates the nature of sediment transfers and storage at the fluvial-coastal interface.

5.5.1 Medium-term change

The Speymouth delta has been one of the most rapidly changing sections of the British coastline over the last two centuries (Grove 1955). The historical record of which was discussed in Section 3.4 (Figure 3.18). The changes are complex, but the dominant tendency is one of westward migration of the river mouth towards Kingston, driven by the westerly construction of spit formations across the mouth. This process can continue uninterrupted for many years and spits of over 5km in length have been documented (Hamilton 1965), although, more frequently the westerly migration of the mouth is checked by natural (or engineered) breaches in the spit (see Table 3.1). These phases of spit growth and breaching have implications for the storage and transfer of fluvial sediment to the coast, which will be investigated herein.

5.5.2 Short-term (contemporary) storage and change

The Speymouth delta is defined herein as the complex series of spits and bars deposited at the mouth of the River Spey which extend ca. 1.5km along the coast from profile -0.1 in the east at Tugnet to profile -1.5 in the west at Kingston (Figure 3.21 and Plate 3.3). These spit formations result from continual shifting of the river mouth (Figure 3.18) and represent the result of the interaction of fluvial and coastal processes. Both the subaerial and submarine extents of the deposits were surveyed in order to quantify the total volume of stored sediment.

Digital elevation models of the subaerial delta (Figure 5.24) were constructed from survey data using the methods set out in Section 4.6.1. Storage volumes were computed using a horizontal base level which lies just below the lowest point of the survey. As the survey extends to the low tide limit on both the seaward and landward side, a base level equal to the elevation of LWMS (-1.4m) was used. The storage volumes presented below thus refer to sediment above LWMS only, although it is recognised that potential stores of sediment and mobile sediment lie beneath this level. The depth of sediment at the delta is difficult to assess and the storage volumes in Table 5.9 are minimum estimates.

In May 1997, 521 000m³ of sediment was stored above LWMS in the Speymouth delta complex (Table 5.9). 60% of this sediment was stored immediately west of the river mouth to profile -0.8, while the eastern side of the delta complex stores only 15% of the total delta sediment. The remaining 25% is stored between profile -0.8 to -1.5. Between May 1997

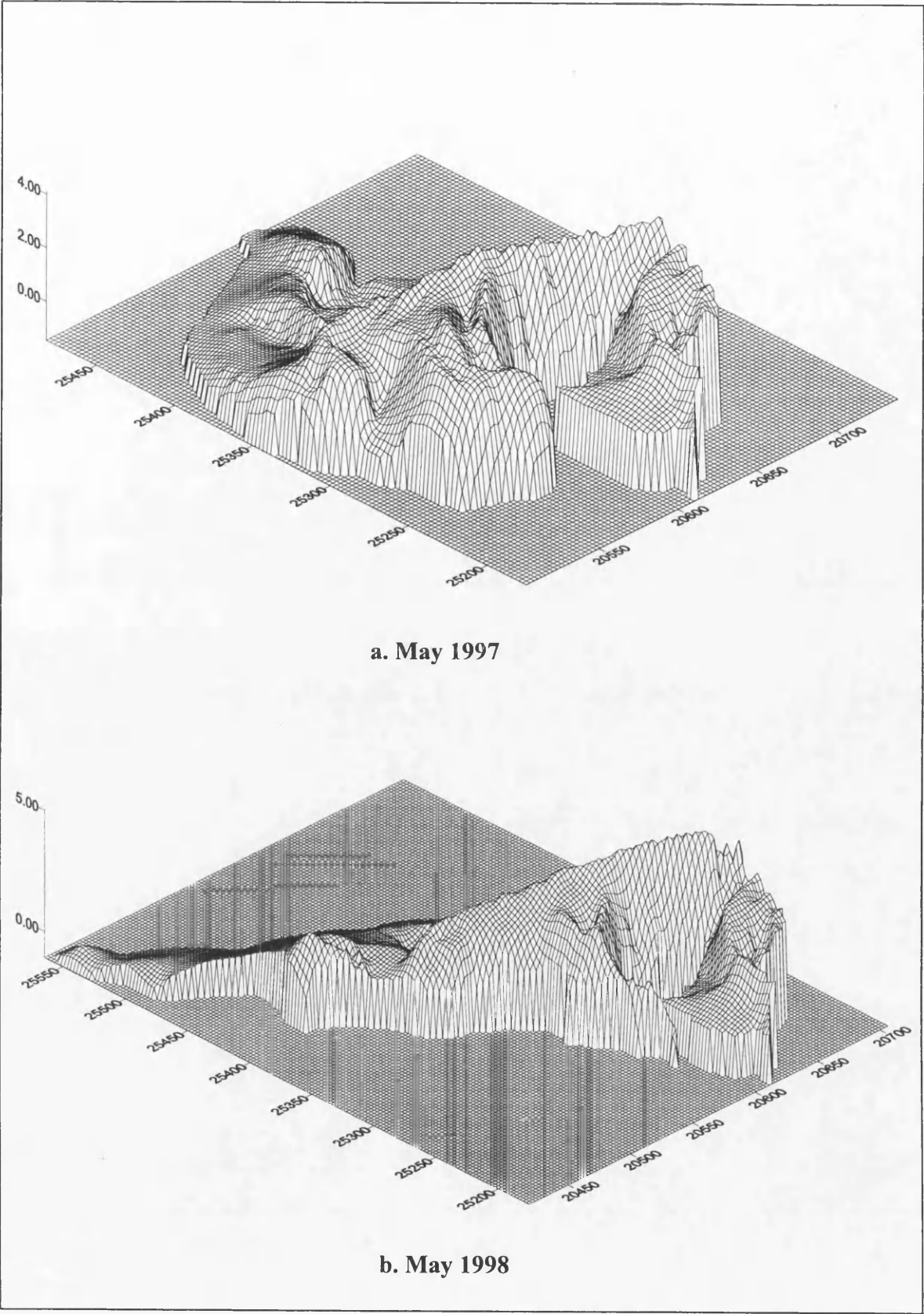


Figure 5.24: Digital elevation models of the subaerial delta at Tugnet, delimited by profile -0.1 a. May 1997 b. May 1998

and May 1998 the subaerial delta gained a total of $4\,314 \pm 173 \text{ m}^3$ of sediment (Table 5.9). The error range is defined on the basis that storage changes of this magnitude are accurate to within $\pm 4\%$, with a survey point spacing of ca. 6m (Figure 4.9a, Section 4.6.1.6).

	Storage volumes (m^3)		Storage change (m^3)
	May 97	May 98	May 97 to May 98
profile -0.1 to river	78 429	84 738	+6 309
river to profile -0.5	129 626	117 309	-12 316
profile -0.5 to -0.8	171 361	181 682	+10 321
profile -0.8 to -1.0	62 157	no data	n.a.
profile -1.0 to -1.5	79 093	no data	n.a.
total / net change	520 666		+4 314

Table 5.9 : Volume of sediment stored at the Speymouth delta in May 1997 and May 1998. For volume calculations the delta is split into sections delimited by beach profiles (locations shown in Figure 3.21).

Throughout the study, a low gravel spit extended westwards from Tugnet diverting the river mouth to the west (Table 5.10 and Plate 3.3). In May 1997 the distal end of the spit was ca. 200m west of profile -0.1; by March 1999 the distal end of the spit had migrated westwards by ca. 260m and lay directly opposite profile -0.5 growing at a rate of approximately 150 m a^{-1} (Table 5.10).

date	length of spit from profile -0.1 (m)	growth rate (m a^{-1})
28-May-97	200	
27-May-98	370	170
3-Mar-99	460	120

Table 5.10: Growth of the gravel spit extending westwards from profile -0.1 at Tugnet, recorded during repeat surveys and field observations.

This westerly spit extension diverted the river mouth west and had consequent effects both at the river exit and downdrift on the western flank of the delta complex. The second delta survey in May 1998 documents these changes (Figure 5.24). Erosion at the outer bank of the river as it enters the coast caused a loss of sediment from the landward side of the Tugnet delta complex. However, this loss was compensated by the gain in sediment due to spit growth. Overall, the eastern flank of the delta increased in volume by ca. $6\,300 \text{ m}^3$ between May 1997 and May 1998 (Table 5.9). Erosion and sediment loss also occurred on the west bank of the river and the lobe of sediment which extended eastwards from the Kingston side of the delta in May 1997 was removed as the river exit migrated westwards. This loss of sediment was partially compensated by the accretion of low recurving gravel ridges to the foreshore opposite profile -0.5 (see Figure 6.5 and the discussion in Section 6.3). These ridges are prominent at low tide, reaching altitudes of up to 2m above the foreshore (Plate 5.2). During repeat field visits these features were observed to migrate

westwards along the lower foreshore and became welded onto the lower beach in the vicinity of profile -0.5. In spite of this accretion, a loss of ca. $-12\,000\text{ m}^3$ of sediment was recorded on the western side of the mouth to profile -0.5 (Table 5.9).

Further west, between profiles -0.5 and -0.8 the delta volume increased by ca. $10\,000\text{ m}^3$ between May 1997 and May 1998 (Table 5.9). This sediment gain is predominately due to the accretion of ridges to the seaward along this section (e.g. Plate 5.2). Downdrift of the accreting ridges towards profile -0.8, erosion and retreat of the active beach crest occurred. However, this retreat does not necessarily cause loss of sediment from the delta as most of the sediment was deposited as large overwash lobes of gravel extending landwards into the lagoon. Erosion was observed from profile -0.8 westwards to profile -1.5 and is summarised by the beach profile data presented in Chapter 6.

Nearshore surveys (below LWMS) of the sub-aqueous delta morphology were carried out twice (August 1998 and August 1999) as shown in Figure 5.25. Both DEMs show a gently sloping foreshore to -6m OD ca. 500m offshore. While there is evidence of some morphological variation in the nearshore, there is little evidence to support the presence of a major sub-aqueous delta feature. It is suggested that the apex of the delta lies ca. 8.2km offshore (Chesher and Lawson 1983, Section 3.5.2.2) and deviation of the 10 and 20m depth contours at the mouth of the Spey provide evidence to support this (see Section 6.4 and Figure 6.10). Minor changes in the nearshore topography were observed between surveys (Figure 5.25) although to accurately quantify these volume changes a much denser survey network is required (cf. Hicks and Hume 1997). Topographic variation in the nearshore may be related to offshore sediment accumulations and bars (see Figure 6.11).

5.5.3 Sediment characteristics

Bulk samples of surface and sub-surface sediment at the subaerial delta are presented in Appendix D. Of the three samples, only the sample at Tugnet contains significant proportions of sand (8%) with the two Kingston samples containing negligible amounts of sand (less than 1%). The mean D_{50} of the surface and sub-surface delta sediment are 44 and 29mm, respectively. The mean D_{84} is 83 and 59mm, respectively. In general, the delta samples contain a wider range of grain sizes than the beach bulk samples (Section 6.5.2), but a much narrower range than the river samples (e.g. compare Figure 5.26 to 5.14 and 6.15). The proportion of sand in the entire subaerial delta deposit is difficult to define as fines are likely to be washed to the base of the deposit. Samples taken from the surface and shallow sub-surface will thus contain lower amounts of sand than is contained in the entire deposit. For the sediment budget calculations in Chapter 7 it is assumed that the subaerial



Plate 5.1: Erosion of the left bank of the Spey at the Kingston and Garmouth golf course (May 1996) (NJ 345646).



Plate 5.2: Recurving gravel ridges accreting on the lower foreshore at Speymouth. Note figure for scale (March 1999) (NJ 344659).

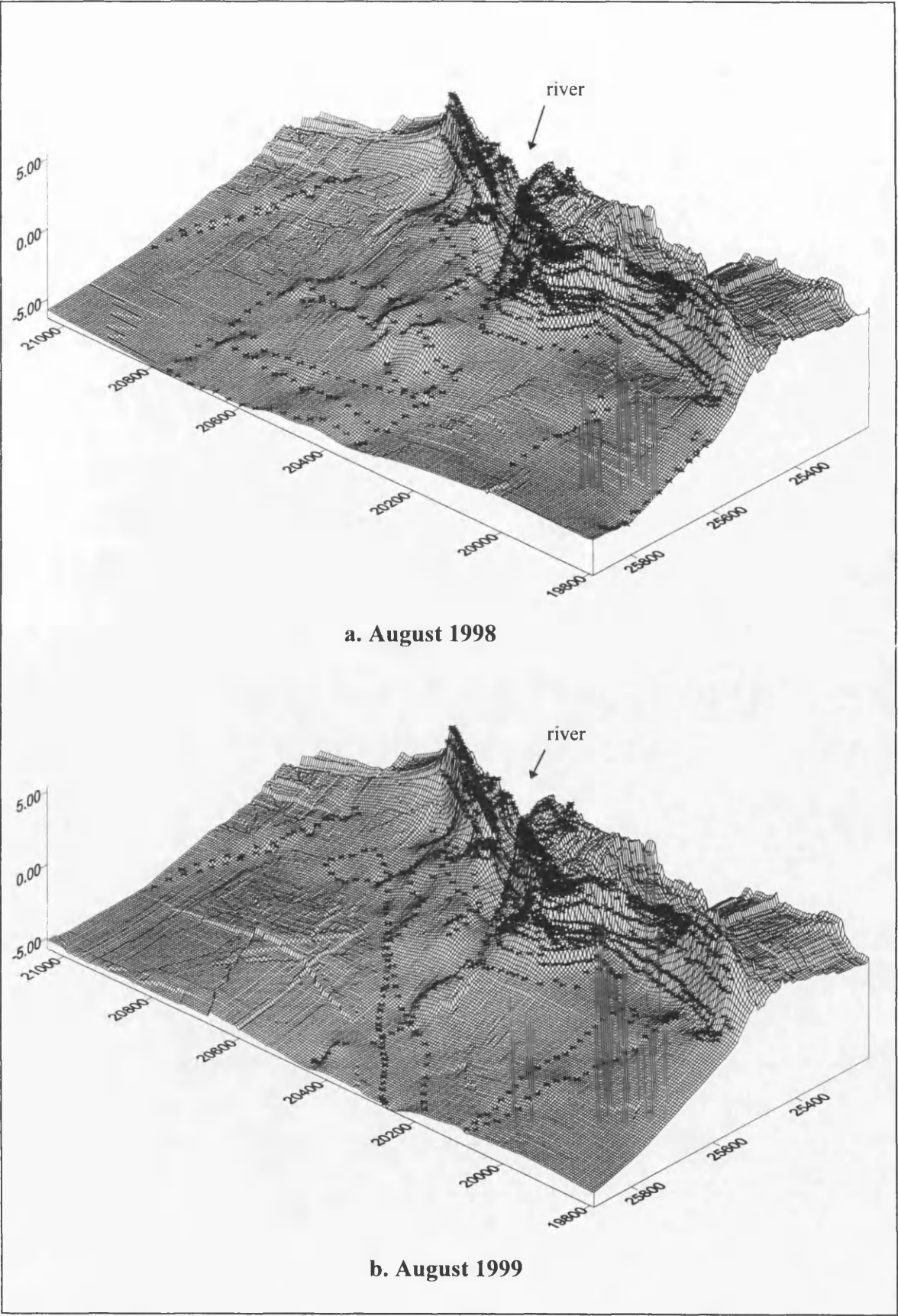


Figure 5.25: Sub-aqueous delta morphology a. August 1998 and b. August 1999. The subaerial delta is as surveyed in May 1998. Crosses represent the survey data.

delta is composed of 5% sand and 95% gravel. The proportion of sand is lower than that of the fluvial input (20% sand, Section 5.3.2) as sand tends to accumulate on the lower beach and nearshore when it enters the coastal system.

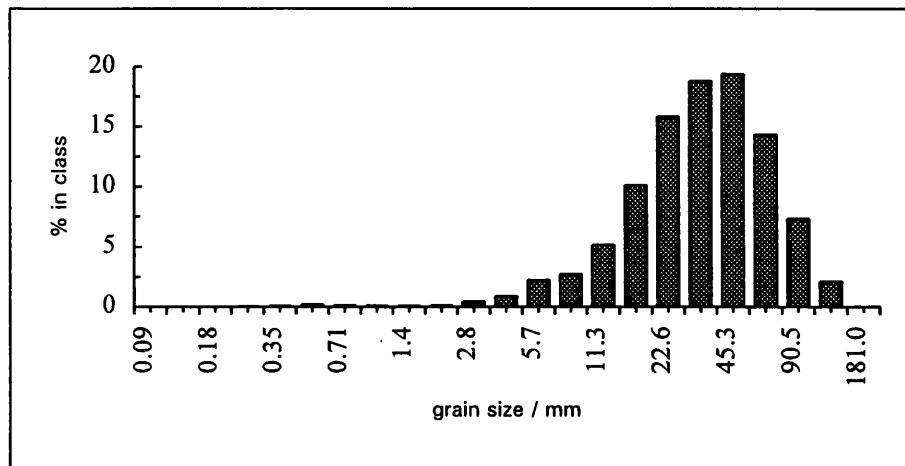


Figure 5.26: Surface grain-size at the Kingston delta (seaward sample) obtained from bulk sampling. Sample size is 978kg. $D_{50} = 40\text{mm}$. Sorting $= \sqrt{(D_{84}/D_{16})} = 2.06$.

Nearshore surface sampling indicates that the nearshore sea-bed at the delta consists of coarse to fine sand (Babtie Dobbie Ltd. 1994), although borehole surveys indicate gravel layers and the presence of occasional rounded medium and coarse gravels within the mainly sandy deposit (Figure 3.24, Section 3.5.2.2). Observations of the nearshore sediment characteristics made during the offshore delta survey in August 1999 indicate that the seabed surface up to 500m offshore is mainly gravel with a patchy sand cover. The river channel bed is gravel, with infilled patches of sand in places. Gravel extends westwards at depth from the river exit to profile -1.5, although the proportion of gravel on the seabed decreases moving westwards from the mouth and is replaced with a sandy veneer. Exposures of seabed gravel visible at low-tide are generally large, algae-covered and well-rounded gravel, indicating a generally immobile bottom sediment. The seabed surface from profile -2.0 westwards is predominately sand, with occasional gravel patches.

5.5.4 Sediment budget implications

The River Spey may transport between 6 000 m³ and 41 000 m³ of sediment annually to the mouth (Section 5.4.2). Between May 1997 and May 1998 ca. 4 000 m³ of sediment accumulated at the delta when fluvial transport rates were ca. 6 000 m³ a⁻¹, therefore storage at the Speymouth delta greatly reduced the amount of fluvial sediment reaching the coast. This may provide an explanation for the high rates of erosion recorded along the coast of Spey Bay during the study period (Chapter 6).

Alternate cycles of spit growth and breaching may have implications for the storage and transfers of sediment from the river to the coast. For example, periods of spit growth results in temporary storage and accumulation of sediment at the delta, while spit breaching injects the stored sediment into the coastal system. A similar process was observed at the Speymouth delta (Figure 5.27) which has important implications for the downdrift coast. The study period was characterised by a period of gravel spit accretion and westerly extension of a spit complex across the mouth of the Spey (Section 5.5.2). The spit extended at a mean rate of 150 m a^{-1} over three years, temporarily storing large quantities of drift sediment and diverting the river exit westwards. As the river migrates westwards fluvial erosion occurs at the western side of the delta, providing an injection of previously stored sediment into the coastal system. Downdrift, an accretionary lobe of sediment is deposited in the lee of the spit, as this part of the coast is sheltered from waves. This sediment is likely to be a combination of fluvial input and sediment eroded from the inner part of the delta deposit. As the spit advances this accretional lobe also advances westwards as a series of recurving ridges which become welded on to the lower foreshore (Plate 5.2). During the first two years of the study the lobe of deposition included profile -0.5, with several new ridges forming to the seaward of the active beach ridge (see Figure 6.5).

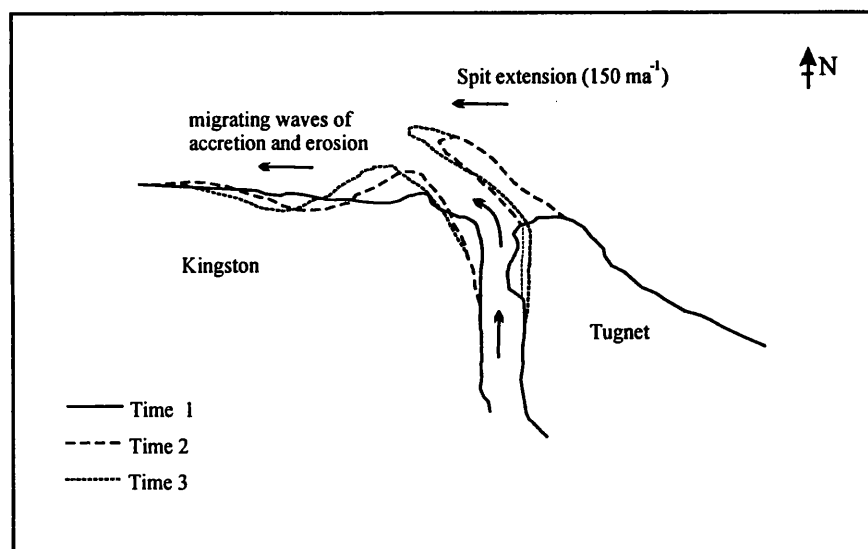


Figure 5.27: Conceptual model of morphological response during a period of westerly spit extension at the Speymouth delta.

The depositional lobe is preceded by an erosional section or bight (profile -0.8 to -1.5) which was also observed to migrate westwards during the study (see Plate 3.3). Beach profile data presented in Chapter 6 corroborates this observation. Erosion occurs here because the depositional lobe at the river mouth acts as a groyne, reducing sediment supply to the downdrift coast and thus enhances erosion. This process results in the slow movement of zones of erosion and deposition along the coast as a function of spit growth

and breaching. Interaction between fluvial and coastal processes at the river mouth will cause significant variations in the longshore coastal transport rate that are felt most close to the mouth but which are then propagated downdrift. This will be explored further in Chapter 7 when the entire sediment budget for the Spey system is presented.

5.6 Summary

Sediment transport rates in the lower River Spey are quantified using a morphological approach and vary from $41\,000\text{ m}^3\text{ a}^{-1}$ (July 1993-July 1994) to $6\,000\text{ m}^3\text{ a}^{-1}$ (December 1997-December 1999). Transport rates are not directly related to flow conditions and other factors such as sediment supply, sediment accessibility and channel stability are important controls. Delivery of this episodic sediment supply to the coast is influenced by the operation of the delta. Cycles of spit growth and breaching cause significant variations in the supply of sediment to the coast.

6. COASTAL SEDIMENT STORAGE AND TRANSFERS

The supply of sediment from the River Spey to the Speymouth delta was quantified in Chapter 5. This chapter presents the results and analysis of planimetric, morphological and sedimentological data pertaining to the Spey Bay coast. From this, the storage and transfers of sediment in the coastal system are quantified. Estimates and associated uncertainties of beach cell volume changes for each ca. 500m wide cell along the coast are presented. These are combined with the fluvial and deltaic volumetric change data in Chapter 7.

6.1 Medium-term shoreline change

In order to place the short-term coastal changes investigated in this study into context, the medium-term trend of shoreline change at Spey Bay was analysed. The position of mean high water springs (MHWS) and mean low water springs (MLWS) on Ordnance Survey maps surveyed in 1870 and 1970 were compared and a map produced showing apparent shoreline change (Figure 6.1, in sleeve). Whilst reservations exist concerning the accuracy of using cartographic sources to determine coastal change (e.g. Carr 1980), it remains that they provide a good order-of-magnitude approximation of the general trend of coastal plan, and thus volumetric, change.

East Spey Bay

The map evidence shows a general trend of erosion in East Spey Bay between 1870 and 1970 (Figure 6.1). The 1970 MHWS and MLWS positions have migrated landwards since 1870 over most of this stretch of coast. Maximum landward migration of MHWS occurs at the extremities of East Spey Bay, with up to 50m of recession in the east at profile +4 and ca. 85m of recession at Tugnet (profile 0). In addition to the landward migration of MHWS and MLWS, there is a general steepening of the foreshore (the distance between MHWS and MLWS is substantially less in 1970 than in 1870 (Figure 6.1)). A short stretch of the central part of East Spey Bay (between profiles +1 and +2) appears anomalous as the 1970 MHWS lies up to 20m seaward of the 1870 MHWS (Figure 6.1), indicating that this short stretch of coastline may have undergone accretion over the 100 years.

West Spey Bay

Erosion is evident west of the River Spey exit between 1870 and 1970, with both the MHWS and MLWS positions migrating landwards by up to 60m along the stretch of coastline fronting the village of Kingston (between profiles -1 and -2), matching

documentary evidence of erosion. This erosional trend continues westwards until a point midway between profile -4 and -4.5, which marks the maximum westerly extent of the coarse-clastic beach in 1870 (Figure 6.1). West of this point up to profile -8.5 (over ca. 4km), the beach gradient is substantially steeper in 1970 and the MHWS position has migrated seaward by up to ca. 120m between 1870 and 1970. Beach steepening and seaward migration of MHWS is associated with the westerly migration of the gravel beach, and its replacement of the sand beach (Figure 6.1). The active gravel beach west of Boar's Head Rock is presently accreting westwards, with gravel initially being deposited at the back of the sand beach but progressively replacing it (Plate 6.1). This westerly accretion of gravel has been prevalent over the last 130 years and is documented by map and photographic evidence (Table 6.1). The total westerly extension of the coarse-clastic beach between 1870 and 1998 is 4270m, an average annual westerly growth rate of ca. 33.3 m a^{-1} , although the growth rate has slowed over time (Table 6.1).

Period	Westerly growth (m)	Growth per annum (m a^{-1})
1870-1903	1360	41
1903-1967	2090	33
1967-1994	720	27
July 1994- Dec. 1995	30	20
Dec. 1995 - Dec. 1998	70	23
TOTAL (1870 - 1998)	4270	33

Table 6.1: Westerly extension of the active coarse-clastic beach (West Spey Bay)

The volume of gravel involved in the extension can be estimated over the 128 years of record. Using a ridge altitude of 4m ASL, a mean beach width of 50 m extending over 2500 m and then tapering to zero over the next 1770 m over 128 years, suggests that over $677\,000 \text{ m}^3$ of gravel has accumulated at $5\,300 \text{ m}^3 \text{ a}^{-1}$. If a 6m closure depth for gravel movement is included to reflect the build-up of gravel below the water level (Comber 1993), then this figure could potentially rise to $13\,000 \text{ m}^3 \text{ a}^{-1}$, although this may be regarded as an absolute upper limit.

The transition from a low-gradient wide sandy beach to a steep coarse-clastic beach over such a relatively short period of time is unusual. For example, in 1870 the beach at profile -8.0 was a 150m wide, gently sloping (ca. 1°) inter-tidal sand beach (Figure 6.1). In 1999 it was a ca. 60m wide coarse-clastic beach, consisting of an steep (ca. 7°) active beach ridge at ca. 6m OD with several gentle landward curving ridges in the backslope at slightly lower altitudes (Plate 6.2 and Appendix F).



Plate 6.1: Westerly extent of the coarse-clastic beach ca. 3km east of Lossiemouth (looking to the east) (April 1996) (NJ 264687).



Plate 6.2: Wide coarse-clastic beach with recurving ridges in the backslope (profile -8) (looking to the west) (NJ 277681). In 1870 the beach here was a ca. 150m wide, gently sloping inter-tidal sand beach beyond the westerly limit of the encroaching gravel.

West of the gravel beach front, map evidence shows the 1970 MHWS position up to 40m further landwards from its 1870 position, indicating that the sand beach is generally erosional along this stretch of coastline to Lossiemouth (Figure 6.1). Frontal dune erosion is evident today along this stretch of coastline (Plate 3.6, Section 3.5.2.1).

6.2 Beach volume change (1870-1970)

Volumetric changes along the coast of Spey Bay were estimated from the changes in beach profiles between 1870 and 1970. A series of beach profiles (Figure 6.2) were drawn every 0.5km along the Spey Bay coast based on the plotted positions of MHWS and MLWS, at locations matching those regularly surveyed in this project (Figure 6.1). Profile 0 is located at Tugnet, positive profiles are distances east of this point and negative profiles are distances west of this point (Figure 3.21). Volume changes between successive profiles (Table 6.2) were calculated using the mean end areas rule (equation 2.11, Section 2.1.2).

There is an increase in beach gradient at most profiles between 1870 and 1970, especially in western Spey Bay (Table 6.2 and Figure 6.2). In some cases volume gains from profile accretion are exceeded by volume loss due to profile steepening. There are difficulties in such analyses since, in addition to cartographic errors, the recession of the coast may not always result in loss of sediment. Where the beach is backed by low lying ground, the beach may simply “roll-over” landwards. Where it is backed by raised beaches, such as at Spey Bay, landwards movement of the coast will involve loss of sediment. Gravel beaches are steeper than sand beaches and where a sand beach is replaced by a gravel beach the former are diminished in area resulting in a loss in volume. In addition, the mapped volumes are almost certainly under-estimates since if the volume of sediment beneath the water level is included then the actual gain of gravel depends crucially on the closure depth of gravel adopted (Section 2.1.2).

Profile	1870 beach gradient	1970 beach gradient	cell volume change /m ³
East Spey Bay			
+4.0	0.02	0.03	- 128 000
+3.5	0.03	0.04	- 126 000
+3.0	0.02	0.07	- 104 000
+2.5	0.02	0.03	- 60 000
+2.0	0.02	0.04	- 45 000
+1.5	0.02	0.05	- 43 000
+1.0	0.02	0.07	- 79 000
+0.5	0.03	0.06	- 145 000
0	0.05	0.12	

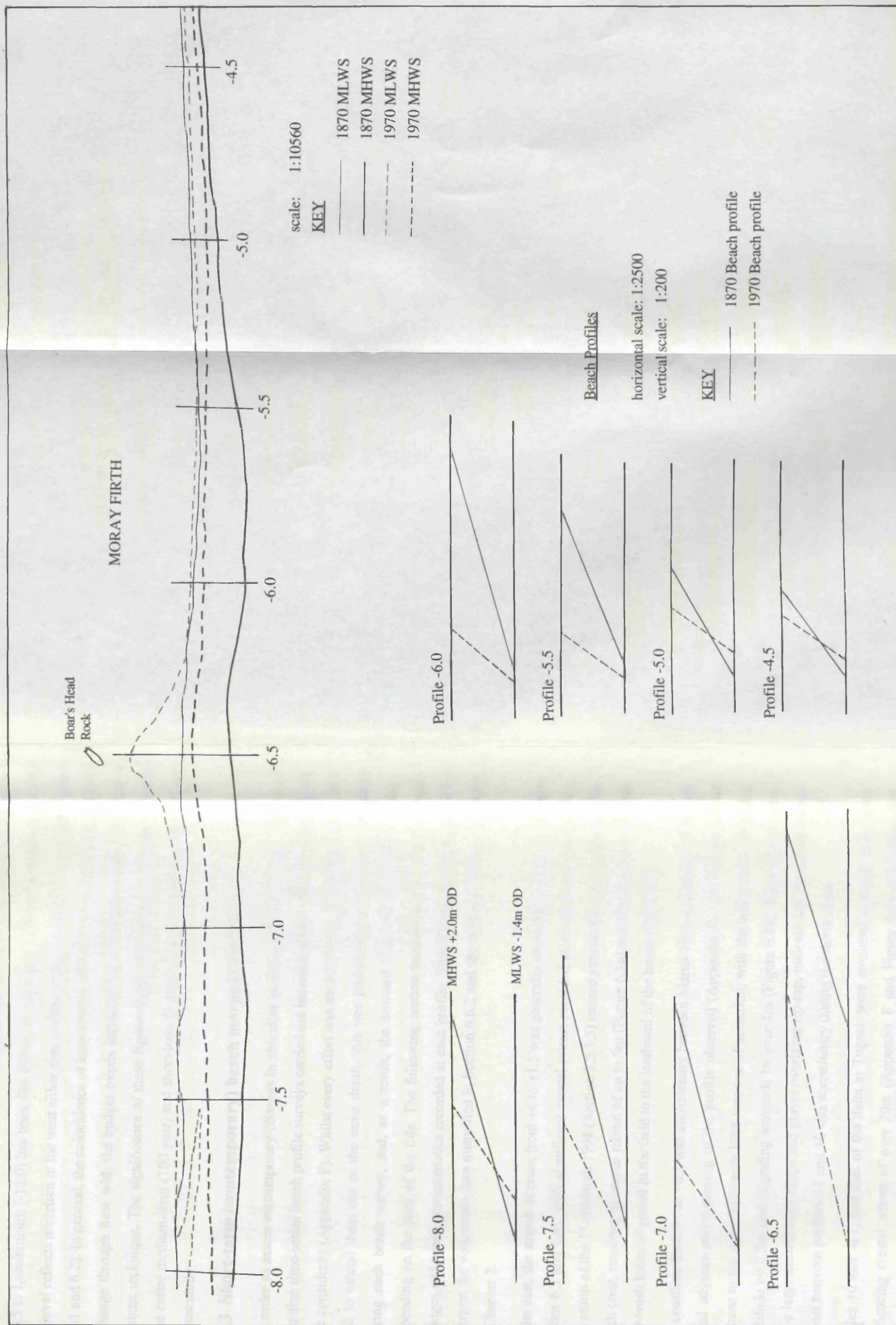


Figure 6.2: Beach profiles in the vicinity of Boar's Head rock plotted from the Ordnance Survey position of MHWS and MLWS, 1870 and 1970 (see Figure 6.1)

authority to the technique. The total loss of sand over the same time period from profile -8.5 to Lossiemouth (-12.0) has been 489 000 m³ or 4890 m³a⁻¹. The replacement of sand by gravel reflects accretion in the west either side of Boar's Head Rock (Table 6.2 and Figures 6.1 and 6.2). In general, the coincidence of known areas of erosion and deposition and their change through time with the budget trends identified by the above mapping suggests a robust technique. The significance of these figures is pursued further in Chapter 7, when the entire medium-term (100 year) and short-term (3 year) Spey Bay sediment budgets are presented.

6.3 Short-term (contemporary) beach morphological change

In order to assess contemporary changes in shoreline position and morphology, results of the nine time-series beach profile surveys carried out between March 1996 and March 1999 are considered (Appendix F). Whilst every effort was made to survey all profiles at MLWS and to extend them out to the same depth, this was problematic given time constraints during each beach survey, and, as a result, the seaward length of each profile varies depending on the state of the tide. The following section summarises the morphological changes and beach characteristics recorded at each profile. This information is then used to interpret the volumetric data quantified in Section 6.6.2 and the sediment budget presented in Chapter 7.

In the east, the stretch of coast from +4 to +1.5 was generally erosional over the three years (Plates 6.3 and 6.4), with significant crestral retreat recorded at all profiles (Appendix F). The storm of the 1st of March 1998 (Section 3.5.1.5) caused substantial over-washing of the beach crest, resulting in crestral retreat of up to 9m (Figure 6.3a) and the deposition of large over-wash lobes of gravel in the field to the landward of the beach (Plate 6.5).

The coastline from +1 to +0.5 was accretionary between March 1996 and March 1999 with crestral advance and flattening of the profile observed (Appendix F). The morphological response to the storm of March 1998 was one of accretion, with the beach crest increasing in altitude by 0.5m and migrating seawards by over 1m (Figure 6.3b). Field observations, such a large accretionary cusps and gravel overlying rip-rap, indicate that the entire stretch of coast between profiles +1 and +0 was accretionary during the storm event.

Profiles +0 and -0.1, just east of the delta at Tugnet, were erosional over the three years experiencing crestral retreat of over 20m (Appendix F and Figure 6.4), suggesting that sediment from this part of the beach may fuel the distal extension of the spit which extended westwards across the mouth of the Spey at mean rate of ca. 150 m a⁻¹ during the

study (Section 5.5.2). Interestingly, the March 1998 storm did not initiate further erosion, but caused an increase in beach crest altitude of ca. 1m (Figure 6.4)..

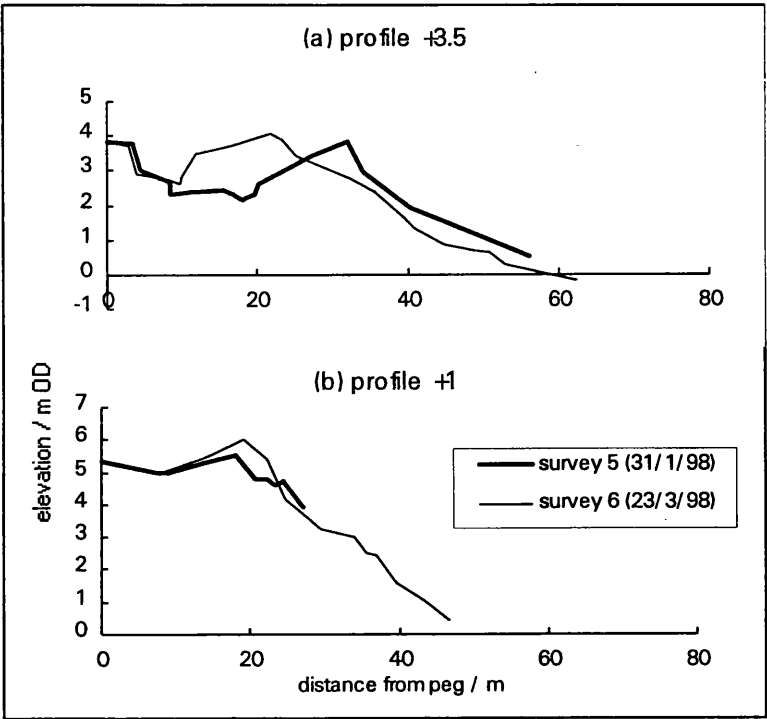


Figure 6.3: Morphological response to the storm of March 1998 at (a) profile +3.5 and (b) profile +1 (ca. 3km downdrift).

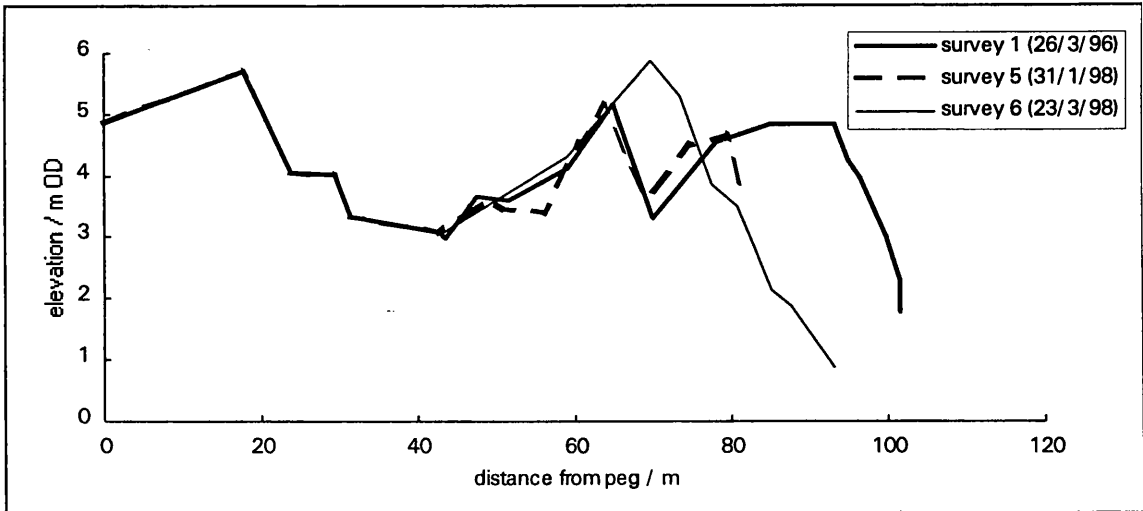


Figure 6.4: Morphological change at profile +0, immediately east of the delta at Tugnet (March 1996, January 1998 and March 1998).

Immediately west of the river mouth, profile -0.5 exhibits beach accretion followed by a later period of erosion, although over the three year period the beach crest moved seawards by ca. 30m (Figure 6.5). The phase of accretion, with low ridges welding themselves onto the lower foreshore at profile -0.5 (surveys 1 to 7) provides evidence of the accretionary lobe of sediment which is deposited in the lee of the advancing westerly spit across the mouth of the Spey (Section 5.5). The trimming of the ridges, observed during surveys 8



Plate 6.3: Profile +4, looking east towards Porttannachy, marks the eastern limit of the study site (NJ 388642). Note erosion of the track at the back of the beach, rubble down the beach face and the low inter-tidal rock platform to seaward (March 1996).



Plate 6.4: Erosion between profile +4 and +3.5 (September 1997)(NJ 385643). Several large erosional scour holes were observed in the back of the beach along this stretch of coast.



Plate 6.5: Large over-wash lobes of gravel extending into the field to the landward of the beach between profile +3.5 and +3 (NJ 385643). Note figure and 4m pole for scale. The beach has retreated by up to 15m along this stretch of coast.



Plate 6.6: Erosion of the World War 2 pill-box, between profiles -1.5 and -2 (NJ 327660). This line of pill-boxes and tank traps extends along the coast of west Spey Bay and would have initially been emplaced landward and parallel to the 1940 shoreline. This indicates long-term erosion of this stretch of coast. The coastal geomorphology map (Figure 3.21) shows the point where the line of tank traps is subsumed by the erosional beach.

and 9, occur as this part of the beach has begun to be affected by fluvial processes as the river mouth migrates further westwards. The time-series of changes at profile -0.5 thus lend support to the fluvial-coastal interactions discussed in Section 5.5. The most seaward ridge included in survey 1 was artificially removed prior to survey 2 to provide recharge sediment emplaced on the downdrift coast (between profiles -1.0 and -1.5).

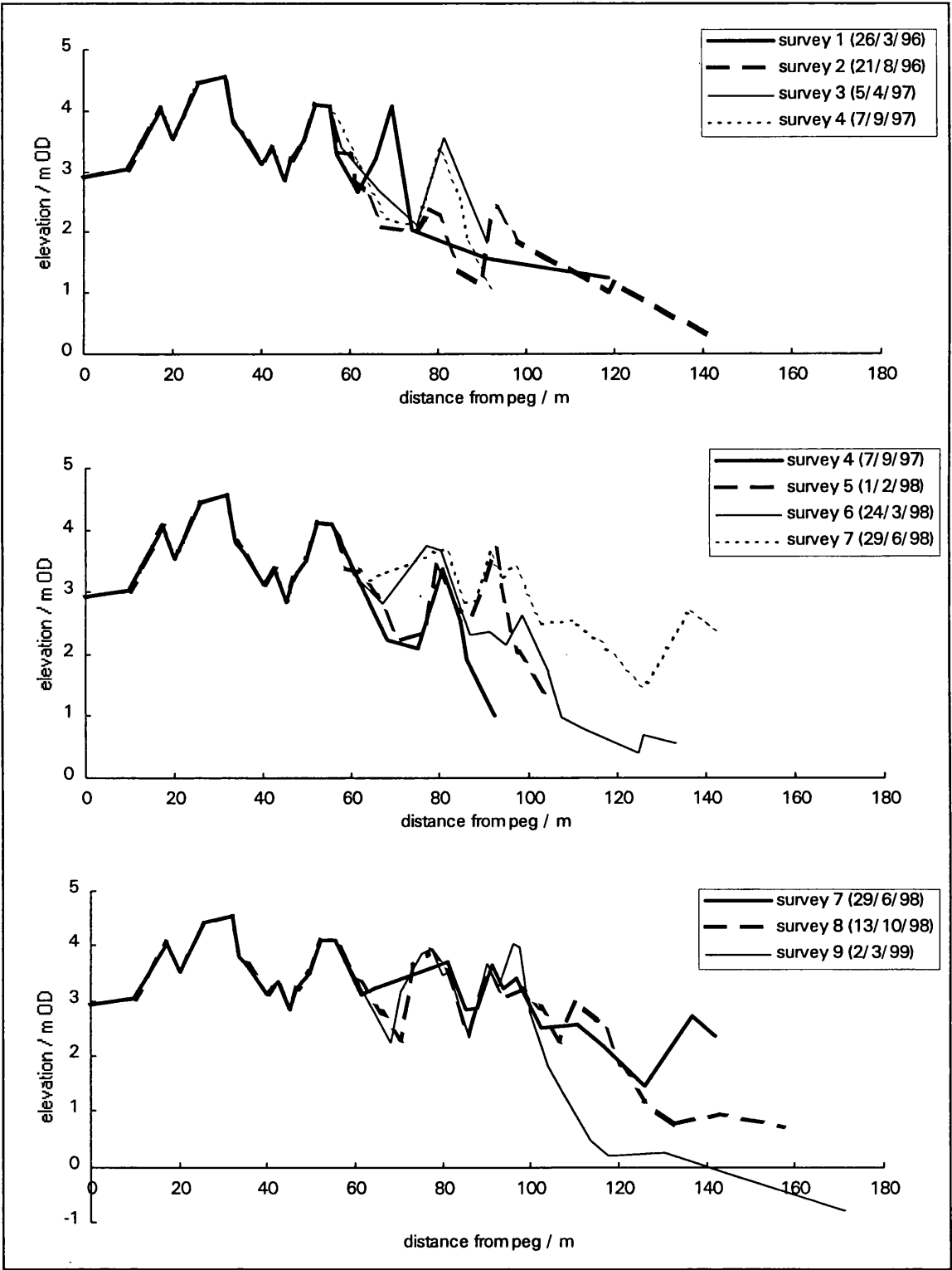


Figure 6.5: Profile -0.5 Morphological Change (March 1996 to March 1999)

Further west, profiles -0.8 to -1.5 were generally erosional, undergoing substantial crestal retreat (up to 20m) over the three years of record (Appendix F). The March 1998 storm caused significant morphologic change along this stretch of coast, with up to 13m of crestal retreat recorded at profile -0.8 (Figure 6.6). Beach-face accretion recorded during survey 2 at profile -1.5 (Appendix F) may be the result of the downdrift transfer of beach recharge sediment emplaced along the updrift coast in late March 1996. Long-term erosion of the stretch of coast between profile -1.5 and -2 is evident from the presence of WWII pill-boxes in the active beach face (Plate 6.6) and map evidence (Section 6.1).

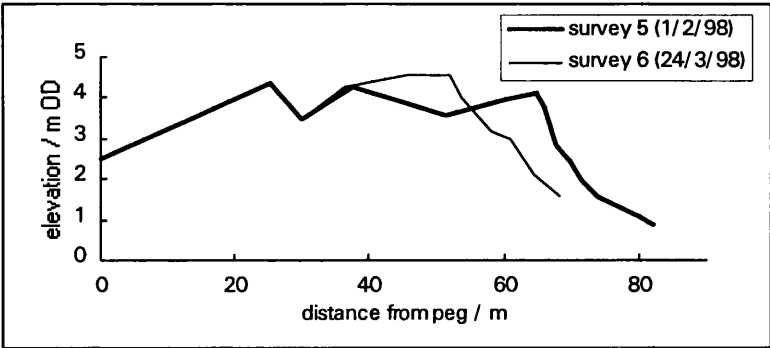


Figure 6.6: Morphological response to the storm of March 1998 at profile -0.8

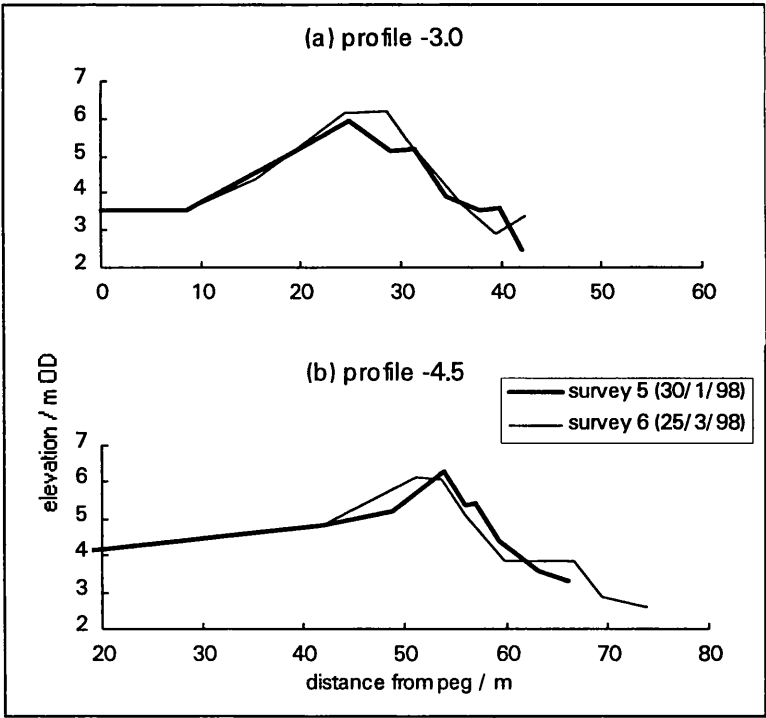


Figure 6.7: Morphological response to the storm of March 1998 at (a) profile -3.0 and (b) profile -4.5 (ca. 1.5km downdrift).

The section of coast from profile -2 to -5.5 recorded only minor morphological changes between March 1996 and March 1999 (Appendix F), suggesting that this part of the coast is relatively stable or is undergoing throughput of gravel, which leaves little or no

morphological signature. The March 1998 storm caused the most significant change along this stretch of coast, with accretion at profiles -3 and -3.5 causing the beach crest to advance seawards by up to 4m and increase slightly in altitude (Figure 6.7a). Further downdrift, at profile -4.5, an erosional response to the storm was recorded, with substantial over-wash and crestral retreat of ca. 3m (Figure 6.7b).

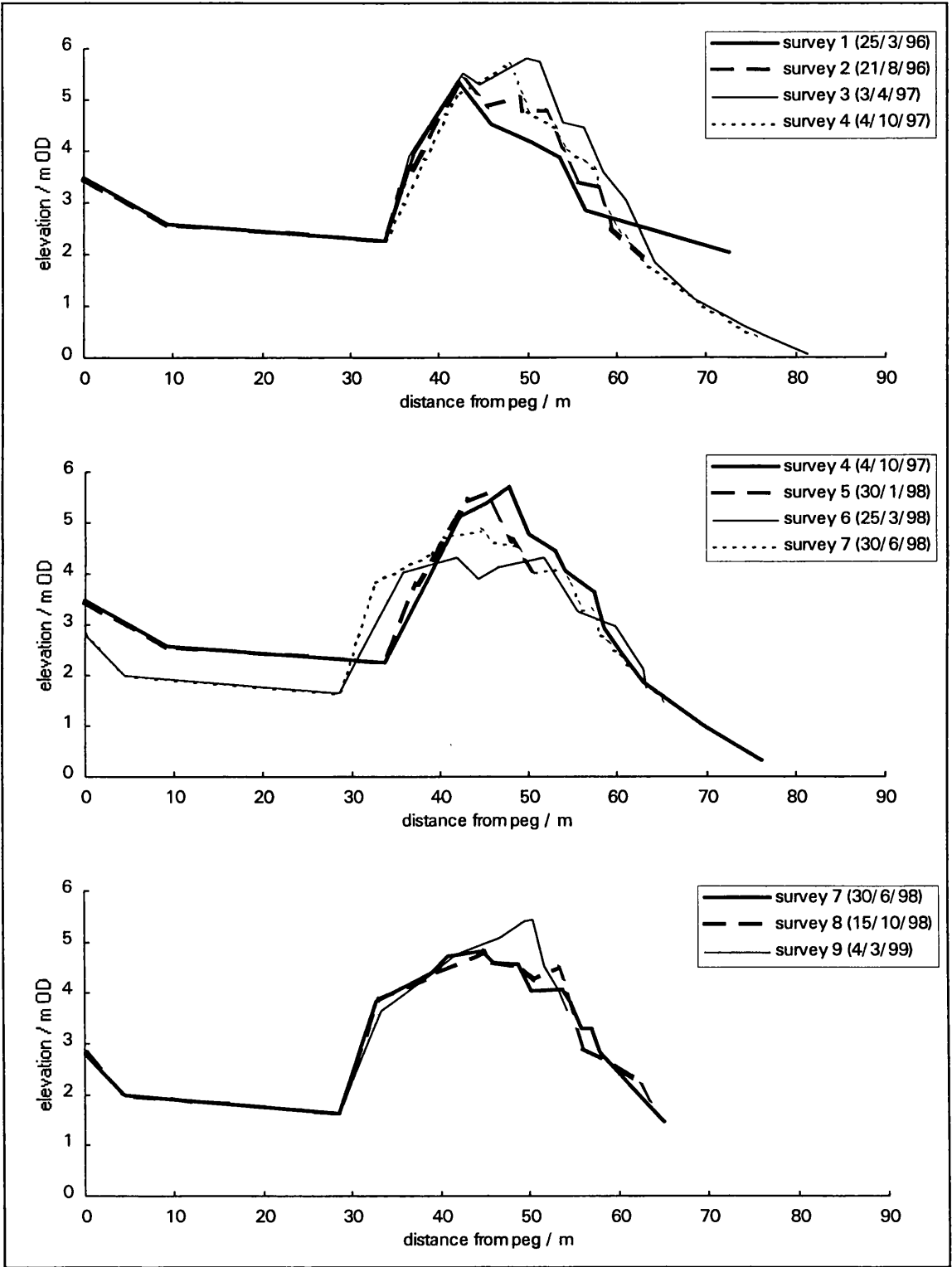


Figure 6.8: Profile -7 Morphological Change (March 1996 to March 1999)

The backslope of the beach at profile -5.5 displays two recurving gravel ridges of lower altitude than the active storm crest (Appendix F and Figure 3.21). The presence of these defined recurving gravel ridges, truncated by the present beach, is thought to represent long term erosion and planimetric readjustment of this part of Spey Bay (Section 3.5.2.1). As profile -5.5 marks the 1870 limit of the westerly migrating gravel beach front (Figure 3.21 and Section 6.1), all gravel west of this point has been deposited since 1870.

To the west, between profiles -6 and -8.5, the beach underwent phases of major accretion and erosion with no dominant trend (Appendix F and Figure 6.8), indicating high amounts of sediment movement and activity. Further west, between profiles -9 to -9.5 only minor changes were recorded in the beach-face (Appendix F).

Profiles -9.75 and -9.8 are backed by high dunes and document the accretion of a gravel beach at the back of the low gradient sand beach (Appendix F). The gravel beach at profile -9.75 increased in width by 22m between September 1996 and October 1998, as the gravel-sand boundary on the lower shoreface moved progressively seaward (at a mean rate of ca. 10 m a^{-1}) (Figure 6.9).

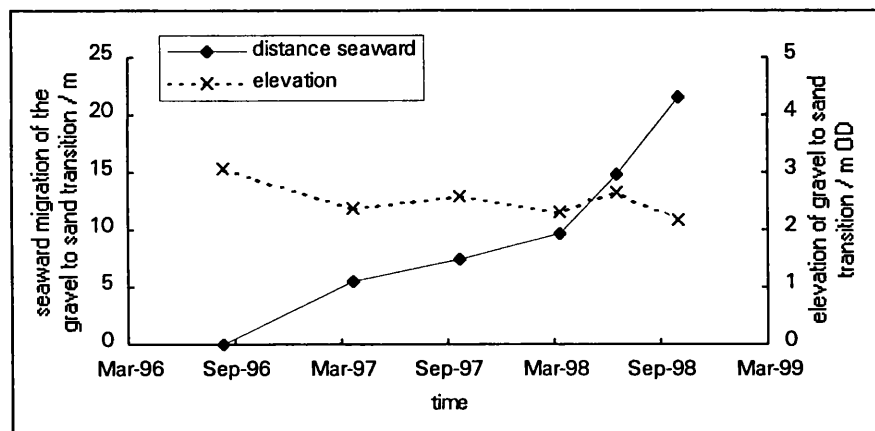


Figure 6.9: Seaward migration and elevation of the gravel to sand transition on the lower shoreface at profile -9.75

Observations at profile -9.8, the extreme westerly gravel limit in March 1996, document the episodic nature of gravel beach migration over the three years. Sand accretion at the base of the dune meant gravel was not visible at the back of the beach again until February 1998 (survey 5). At this time the gravel beach was 7m wide and the extreme westerly limit of gravel lay ca. 25-30m further west. Following the March 1998 storm, the gravel beach was no longer defined and only a scattering of loose gravel remained on an erosional sand beach. By July 1998 a defined, 25m wide, gravel beach draped in a blown sand cover, was present (Plate 6.7) and the extreme westerly limit of gravel lay 75m to the west. By survey 8, the ca. 50m wide upper beach consisted of a mix of sand and gravel and the westerly



Plate 6.7: Beach profile -9.8 (July 1998) (NJ 264687). Note the gravel beach landward of the survey equipment. The gravel beach is overlain by blown sand, but its morphology is distinct with a defined beach crest and cusp features. The instrument position lies ca. 3m seaward of the gravel-sand transition, although there is loose gravel overlying sand further seaward.



Plate 6.8: Beach morphology at profile -2.5 (August 1996) (NJ 326661). Note the cross-shore beach sediment sorting and the accretionary nature of the upper beach with many ridges at different levels. This stretch of beach is undergoing erosion in the medium-term (Section 6.1 and 7.1).

limit lay ca. 70-80m to the west of profile -9.8. During survey 9, sand accretion at the base of the dune had buried the landward part of the gravel beach. At this time, the gravel beach extended 20m west of the profile, although it is likely that further extension was obscured by sand accretion.

West of the gravel beach limit, the section of coast from profiles -10.5 to -12 record the changing morphology of a sand beach backed by dunes (Appendix F). Beach-face slopes are much less steep (ca. 2°) than the gravel beaches to the east (ca. 10°). This part of the beach underwent significant morphological change, particularly where over-washing and/or embryo dune accretion occurred (e.g. profiles -11 and -11.5) and dune erosion (e.g. profile -12).

6.4 Nearshore bathymetry

The bathymetric map extract (based on Admiralty surveys 1898-1918) indicates a gently shelving nearshore at Spey Bay, with nearshore depth contours running approximately shore-parallel up to depths of 5m (Figure 6.10). The 10 and 20m depth contours deviate further offshore at the mouth of the River Spey, which suggests the presence of an offshore delta (see Section 3.5.2.2). Water depths in the map are reduced to Chart Datum, which is approximately the level of Lowest Astronomical Tide and 2.1m below Ordnance Datum (Newlyn). The field bathymetric survey extends each beach profile ca. 500m offshore from HWM (Figure 6.11 and Appendix G). The sea-bed topography is in m OD (Newlyn) so that offshore surveys can be directly related to onshore beach surveys.

East Spey Bay

Offshore extensions of profiles along the coast of east Spey Bay show a very gently sloping nearshore, with depths of only around -3m OD recorded at distances of ca. 300m from HWM. The eastern-most profiles (+4 to +2.5) exhibit very little topographic variation indicating the lack of any constructional features in the nearshore (Figure 6.11a). Further west, and particularly closer to the river mouth, there is increased topographic variation in the seabed, providing evidence of constructional forms (e.g. profile 0 has two offshore bars at distances of ca. 170 and 240m seaward of HWM, Figure 6.11b).

West Spey Bay

Profile -0.5 lies immediately to the west of the mouth of the Spey and the nearshore topography shows the constructional bars just offshore of the mouth (Figure 6.11c). It is likely that this topography changes frequently as new bars are constructed and destroyed.

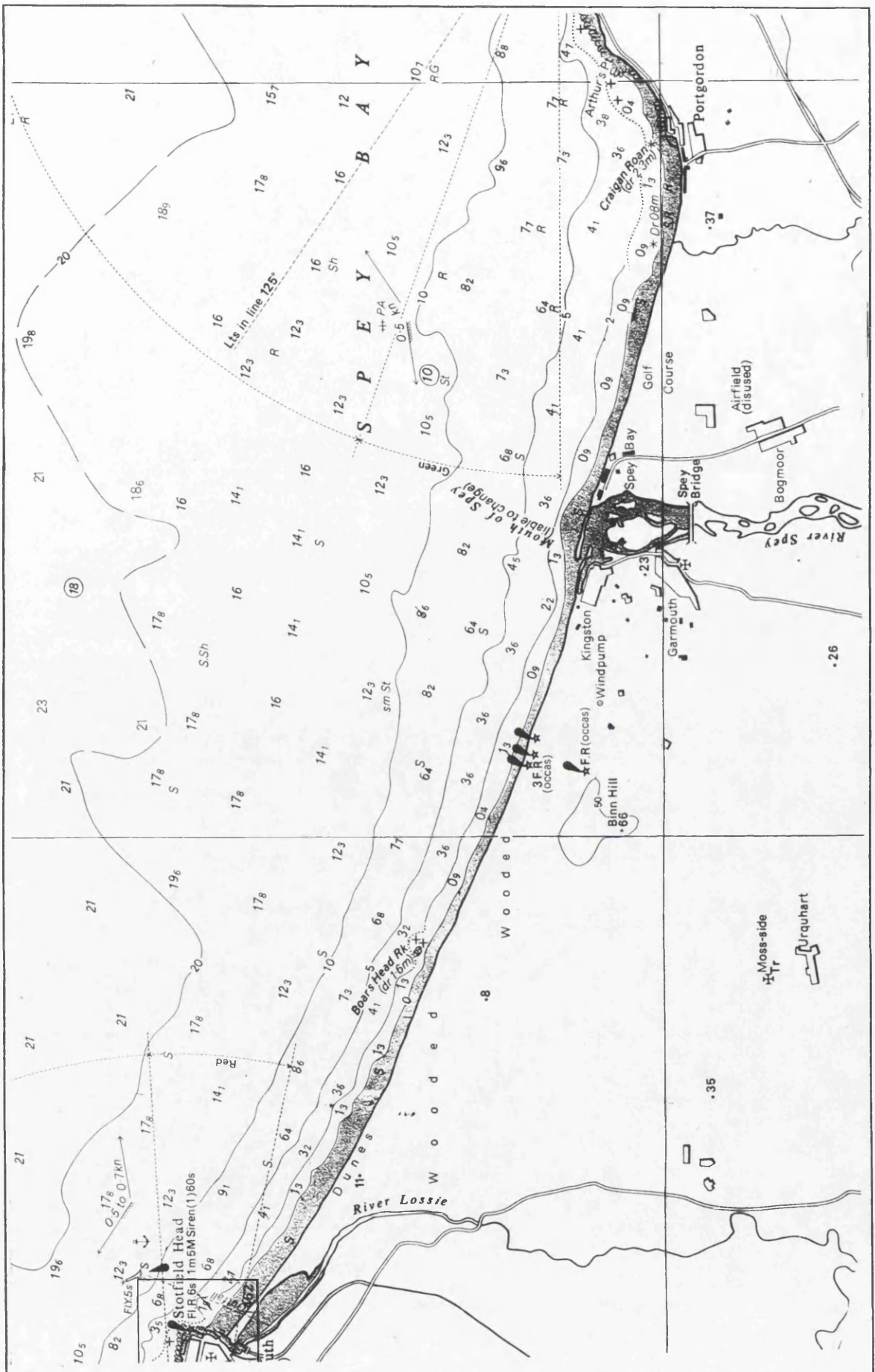


Figure 6.10: Bathymetric Map of Spey Bay (extract from Admiralty chart 223, Dunrobin Point to Buckie). Depths are in metres and are measured to chart datum which is 2.1m below OD (Newlyn). Scale 1: 75 000.

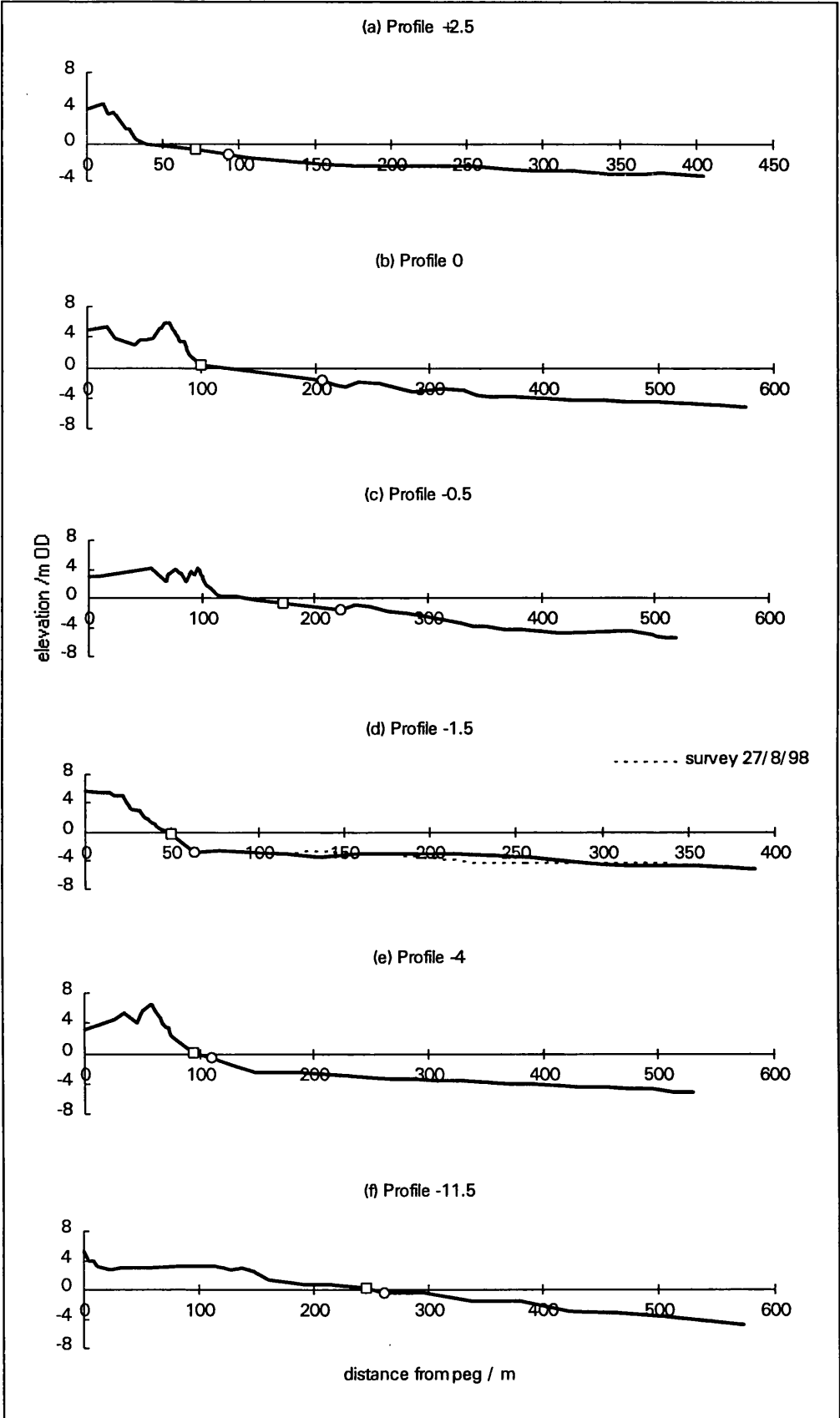


Figure 6.11: Spey Bay beach profiles extended offshore. Boxes and circles show the limits of the onshore and offshore surveys, respectively (see Appendix G for more results).

Profiles -0.8, -1 and -1.5 are the only three to have been surveyed on more than one occasion (August 1998 and 1999) and provide an indication of the nature of annual

nearshore changes along the coast of Spey Bay (Appendix G). From the available data, it appears that there has been negligible change in the nearshore profiles at -0.8 and -1 between surveys. This is in contrast with the repeat survey of profile -1.5 which shows substantial morphological change (and hence sediment transfer) in the nearshore (Figure 6.11d).

A characteristic of the coarse-clastic beach profiles west of profile -1.5 is the steep slope in the immediate nearshore (e.g. profile -4, Figure 6.11e) which tends to flatten out at around -2m OD at a much more gentle gradient offshore. This steep initial slope was noted at all profiles and may be indicative of the transition from a gravel/sandy slope to the gently sloping sandy bottom. Again the nearshore zone is relatively shallow with depths of around -5m OD recorded ca. 450m seaward of HWM. There is little evidence of constructional features in the nearshore along most of this stretch of coast, with the exception of profiles -2 to -4, where two nearshore bars can be traced alongshore at distances of ca. 150 and 270m seaward of HWM (Appendix G).

The offshore profile extensions of the sand beach at Lossiemouth (profiles -9.8 to -12) have a gently sloping nearshore, at approximately the same gradient as the lower beach and do not display the initially steeper profile characteristic of the coarse-clastic beaches (Figure 6.11f). There is evidence of topographical variation in many of the offshore profiles (e.g. profile -11.5) indicating the presence of constructional features such as sand bars. Beach lowering has occurred at some profiles, particularly those towards the western end of the sand beach, with profile -12 experiencing ca. 0.5m lowering between October 1998 and August 1999 (Appendix G). This erosional trend was also observed from comparison of the onshore morphological surveys (Section 6.3).

6.5 Sediment characteristics

The median grain size of the surface beach gravel at Spey Bay varies from 30 to 50mm along the beach, with a mean of 38mm (-5.24ϕ), and shows no obvious downdrift trend, until the abrupt transition from gravel to sand ($D_{50} = 0.22\text{mm}$) ca. 3km from Lossiemouth (Figure 3.22, Section 3.5.2.2).

In order to convert the volumetric transfers of sediment quantified in this study into transfers of gravel and sand, an estimate of the entire particle size distribution of the coastal sediment is required (Section 4.4). In addition, it is important to assess the cross-shore characteristics of the beach sediments, as different events may result in a different calibre

of sediment being transferred (e.g. erosion of the lower beach may involve larger transfers of finer sediment). This section presents the detailed results of beach sediment analysis.

6.5.1 Longshore and cross-shore surface characteristics

The results of all beach surface sediment samples are presented in Appendix H. Cross-shore characteristics can be shown by comparing results from samples taken on the upper, middle and lower beach east of profile -3 (Figure 6.12).

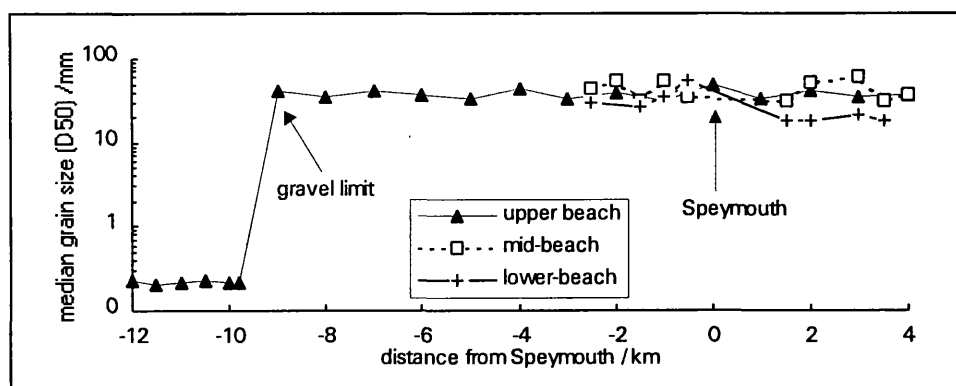


Figure 6.12: Median grain size (D_{50}) in mm of samples taken on the upper, mid and lower beach. West of profile -9.8 all upper beach samples are sand.

The abrupt transition from a gravel to sand beach at profile -9.8 is highlighted. The median grain sizes (D_{50}) of the mid-beach gravel samples are more variable than those of the upper beach (Figure 6.12). This may be a result of the actual sample location on the mid-beach, as it is this part of the beach that is likely to display cusp and berm bed-forms, which have implications for sediment size and shape sorting (Bluck 1999 and Section 2.2.3).

Finer gravel is generally present on the lower beach (mean $D_{50} = 28\text{mm}$) with the exception of profile -0.5 ($D_{50} = 56\text{mm}$) (Figure 6.12 and Appendix H). At the majority of profiles there is a distinct gravel-sand transition on the lower foreshore at ca. LWST, although this lower sand beach is often overlain by scattered fine gravel.

There is little longshore variation in beach sediment sorting (Figure 6.13) and the upper and mid-beach samples show a similar degree of sorting (σ_g ranges from 1.3 to 1.7) indicating a moderately well sorted sediment. Sorting on the lower beach is more variable, ranging from 1.3 to 2.1 (Figure 6.13). The samples at profiles -0.5 and profile -1, just downdrift of the river mouth, display the poorest sorting. The sand beach west of profile -9.8 consists of a relatively well sorted ($\sigma_g = 1.3$) medium sand.

Visual observations of beach sediment show a degree of down profile fining (Plate 6.8), with larger clasts at the crest of the storm beach, fining down beach to fine gravel and sand on the lower foreshore. This sorting is often complicated by complex local sorting

associated with the development of cusp and berm bed-forms, which generate clast assemblages of characteristic grain shape and size at different positions on the beach face (Bluck 1999).

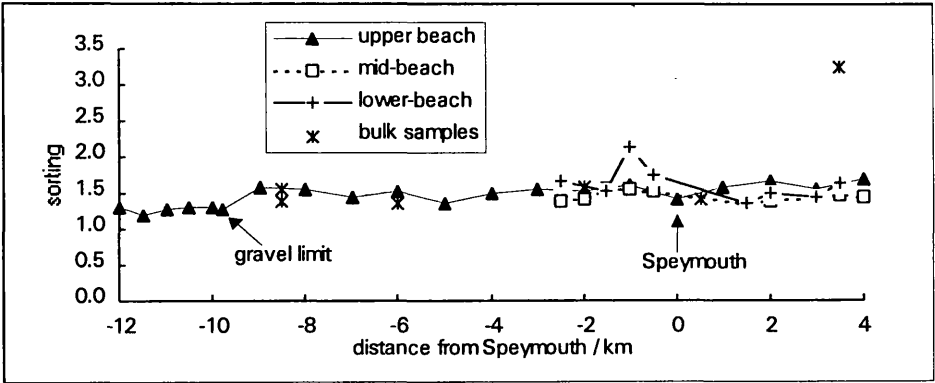


Figure 6.13: Beach sorting of surface samples collected on the upper, middle and lower beach of Spey Bay. The sorting recorded for bulk samples are also plotted. Sorting is defined as $\sigma_g = (D_{84}/D_{16})^{0.5}$.

Skewness is variable both alongshore and cross shore (Figure 6.14), although the majority of results lie within the range +0.1 to -0.1, indicating a near symmetrical distribution. Samples from the mid and lower beach tend to show a slight positive skew, indicating an excess of fines. The sand beach samples (west of profile -9.8) also show a positive skew.

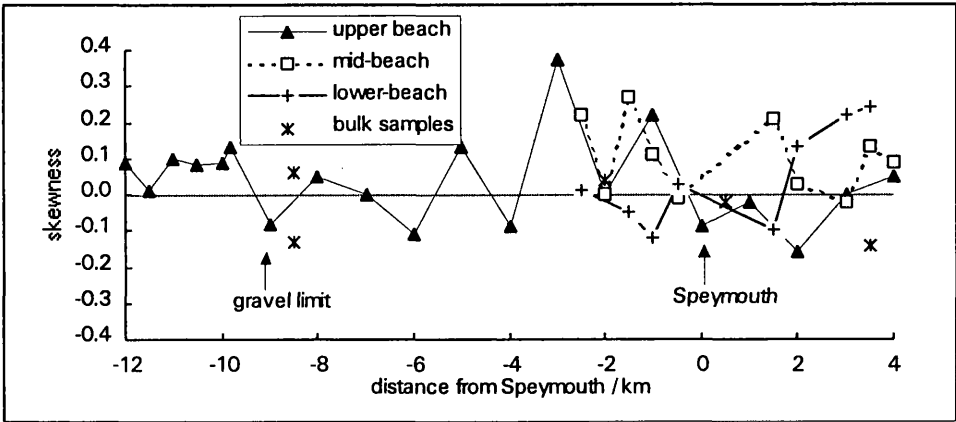


Figure 6.14: Skewness of beach samples collected along the shore of Spey Bay where skewness is defined as $Sk = \frac{\phi_{16} + \phi_{84} - 2(\phi_{50})}{2(\phi_{84} - \phi_{16})} + \frac{\phi_5 + \phi_{95} - 2(\phi_{50})}{2(\phi_{95} - \phi_5)}$.

6.5.2 Bulk samples

Bulk sediment samples were extracted from pits at the top of the main beach ridge (profiles +3.5, +0.5, -6.0 and -8.5) and at approximately HWM (-2.0 and -8.5), adhering to the criteria set out in Section 4.4. Results are presented in Appendix D.

Profile +3.5 is the only gravel beach bulk sample to contain sand, with 8% of the total sample mass less than 2mm. Profile +3.5 is a coarse-clastic barrier beach at the mouth of

Tynet burn which undergoes continual reworking and adjustment in response to both coastal and fluvial processes. The large proportion of fines in this beach sample is likely to be due a combination of the infiltration of fine material as the Tynet burn percolates through the gravel barrier and the relative youth of this part of the beach which was reconstructed less than a year prior to sampling.

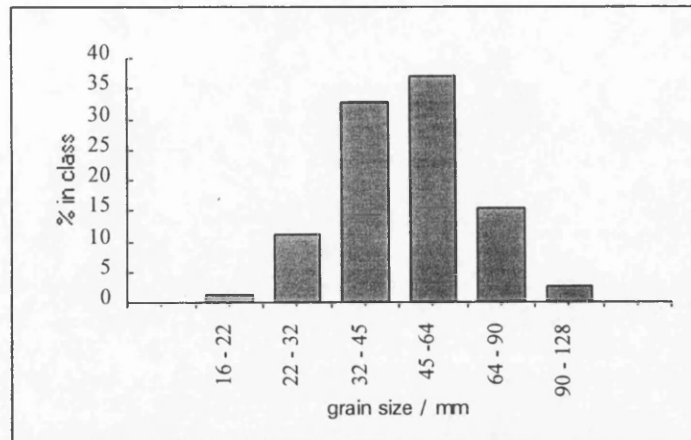


Figure 6.15: Beach sediment characteristics at profile +0.5 (bulk sample size 225kg). $D_{50}=47.4\text{mm}$, $D_{84}=66.8\text{mm}$, sorting $=\sqrt{(D_{84}/D_{16})} = 1.42$.

The beach sediment bulk samples, with the exception of +3.5, are well-sorted and lie within a similar sorting range to that recorded on the upper beach (Figure 6.13 and 6.15). Bulk sample skewness lies within the same range as the surface samples and generally indicate near symmetrical distributions (Figure 6.14).

The D_{50} and D_{84} of the bulk samples show no obvious downdrift trend (Figure 6.16). The samples collected at HWM are generally finer than those extracted from the main ridge crest (Figure 6.16 and Appendix D), as would be expected given the characteristic clast shape and size sorting on gravel beaches (Bluck 1967, 1999).

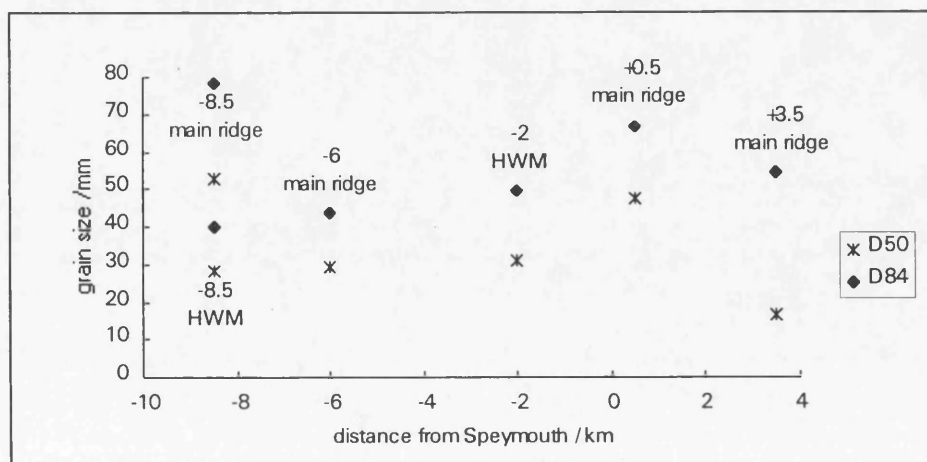


Figure 6.16: D_{50} and D_{84} of Spey Bay beach sediment, obtained from bulk samples of sediment at the locations shown.

While these bulk samples provide a statistically significant representation of the grain size distribution at each sample location, they may not necessarily provide a realistic representation of the relative proportions of sand and gravel involved in the volumetric transfers quantified in Section 6.6. As a result of percolation of fines and size-sorting on the upper beach, the distributions of the samples may not be representative of the entire beach, which is made up of a complex series of clast assemblages, containing different proportions of sand and gravel (Bluck 1999). A representative sample of the proportions of sand and gravel within an entire beach unit is virtually impossible to achieve, given the complexities of beach structure and sedimentation (Bluck 1999).

Based on all the available evidence collected from river, deltaic and beach sediment analysis, it is assumed that sand ($d < 2\text{mm}$) comprises less than 5% of the beach face sediment involved in sediment transfers on the coarse-clastic beach (profiles +4 to -9.75). West of the sand to gravel transition at profile -9.8, it can be assumed that all quantified sediment transfers involve only the sand fraction.

Sediment data collated from borehole samples and nearshore surface samples (Section 3.5.2.2) indicate that the nearshore sea bed sediment is predominately sand with occasional rounded medium to coarse gravels.

6.6 Short-term (contemporary) budget

6.6.1 Closure depth for beach volume calculations

To accurately compare spatial and temporal changes in beach volumes all beach profiles should ideally be the same length and extend to at least the closure depth (see Section 2.1.2). However, as is the case in the majority of studies, the repeat beach profile surveys herein extend to different seaward limits, depending on tidal conditions at the time of re-survey (Section 6.3). This creates problems for both spatial and temporal comparisons between profiles.

Three main approaches to the resolution of these problems can be adopted:

- (1) only the lengths of profiles that have been repeatedly surveyed on *all* occasions are compared. This means that the shortest profile surveyed (usually closest to the time of high tide) defines the limit of beach change that is analysed. Adoption of this strategy means that valuable survey data is ignored and the resulting data-set is depleted.
- (2) comparison of profiles to their surveyed limits. However, this can create erroneous results when comparing temporal variations in profiles, particularly on the lower beach.

For example, if the upper beach experiences cut while the lower beach experiences a compensatory fill between surveys and the limit of the repeat survey details the cut part of the profile *only*, the volumetric result will indicate an erroneously high loss of sediment.

- (3) extend all beach profiles to the same limit (ideally the depth of closure) to create a uniform boundary condition for comparison of changes in volume. Lacey and Peck (1998) extend their profiles to MLWS for volume comparisons. The depth of closure can be defined in numerous ways (see Section 2.1.2), perhaps the most useful being the depth at which there is negligible morphological change in the offshore profile (cf. Jimenez and Sanchez Arcilla 1993). This was the original aim but weather conditions prevented the acquisition of reliable offshore profiles on a time-series basis. Only one reliable survey was possible and so comparison is impossible. Alternatively, the seaward limit for beach volume calculations could be defined by the gravel-sand transition on the lower shoreface, assuming that this marks the approximate seaward limit of active gravel transport (i.e. the depth of closure). However, the limit of this transition varies considerably over time at each profile. A viable alternative is to use the seaward limit or the maximum depth surveyed at each profile for calculations (e.g. Figure 6.17).

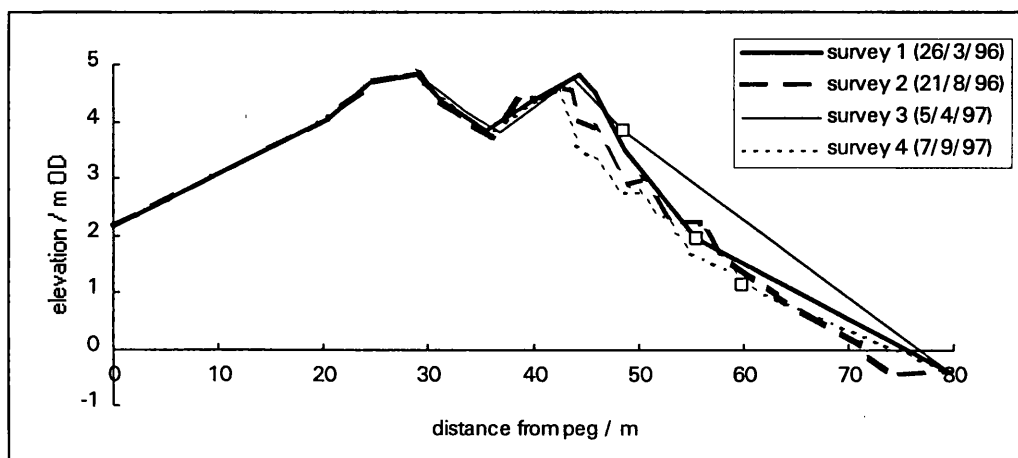


Figure 6.17: Profile -1 extrapolated to the maximum depth surveyed (survey 2). The boxes represent the extent of the actual survey data. Note the potential errors in the estimation of accretion or erosion volumes on the lower beach (e.g. survey 3).

Each repeat survey of the profile was extended to the maximum depth surveyed (Figure 6.17) to allow spatial and temporal comparisons of beach profiles to a common seaward limit. However, errors can result from excessive extrapolation of some profiles to this depth. For example, if the limit of the surveyed profile is high in elevation, representing a ridge, and the profile is extrapolated in a straight line to the defined seaward limit, an

erroneously high area of sediment accumulation will be recorded. Similarly, if the last surveyed point happens to be at the base of a ridge this technique will suggest a large amount of beach erosion.

In this study, volumetric change between surveys was estimated first using approach 2 and then using approach 3. This allows the errors created by extrapolation to a defined seaward limit (approach 3) to be assessed and placed into context. Approach 2 represents the actual volumetric changes that occurred between surveys to the limit of the shorter profile only, and thus may discount potential compensatory cut or fill that may have occurred in the lower non-surveyed part of the beach. On the other hand, extrapolating profiles to a 'depth of closure' can lead to assumptions of large areas of erosion or deposition (Figure 6.17). Results using both approaches are presented below.

6.6.2 Quantification of the coastal sediment budget

6.6.2.1 Temporal patterns of beach profile volume changes

This section presents temporal volumetric changes (in m^3/m) recorded at each beach profile between March 1996 and March 1999 (Figure 6.18). Volumetric results using both the surveyed (approach 2) and extrapolated (approach 3) beach profiles are presented together for comparison. Generally, the volumetric changes calculated using the extrapolated profiles are of larger magnitude than those calculated using the surveyed profile data only. This is to be expected given that the extrapolated comparison will always cover a larger zone than the surveyed comparison (e.g. Figure 6.17). What is of concern is that in a few cases the volumetric change calculated is not only of greater magnitude, but also in the opposite direction (e.g. profile -0.1, August 1996 to April 1997; profile -6.5, January 1998 to March 1998), although when reversals occur, they usually involve only small net volumes. This highlights the sensitivity of the volume calculations to the limits of the survey data. From inspection of the profile data (Appendix F), it can be seen that the extrapolated profile for -0.1 may represent the more realistic estimate of the volumetric change, whereas for -6.5, due to the high elevation of the last survey point on survey 5, the extrapolated profile may significantly over-estimate the amount of lower beach erosion between surveys and the volumetric estimate based on the actual data may be more realistic. This raises issues concerning the reliability of other beach sediment budgets based on profiles which have been extrapolated to a common 'closure depth'.

All profiles experience alternating periods of erosion and accretion over the three year study period (Figure 6.18) with no profile undergoing continuous erosion or accretion.

Figure 6.18: Volume changes at each profile in each survey period (in m³/m)

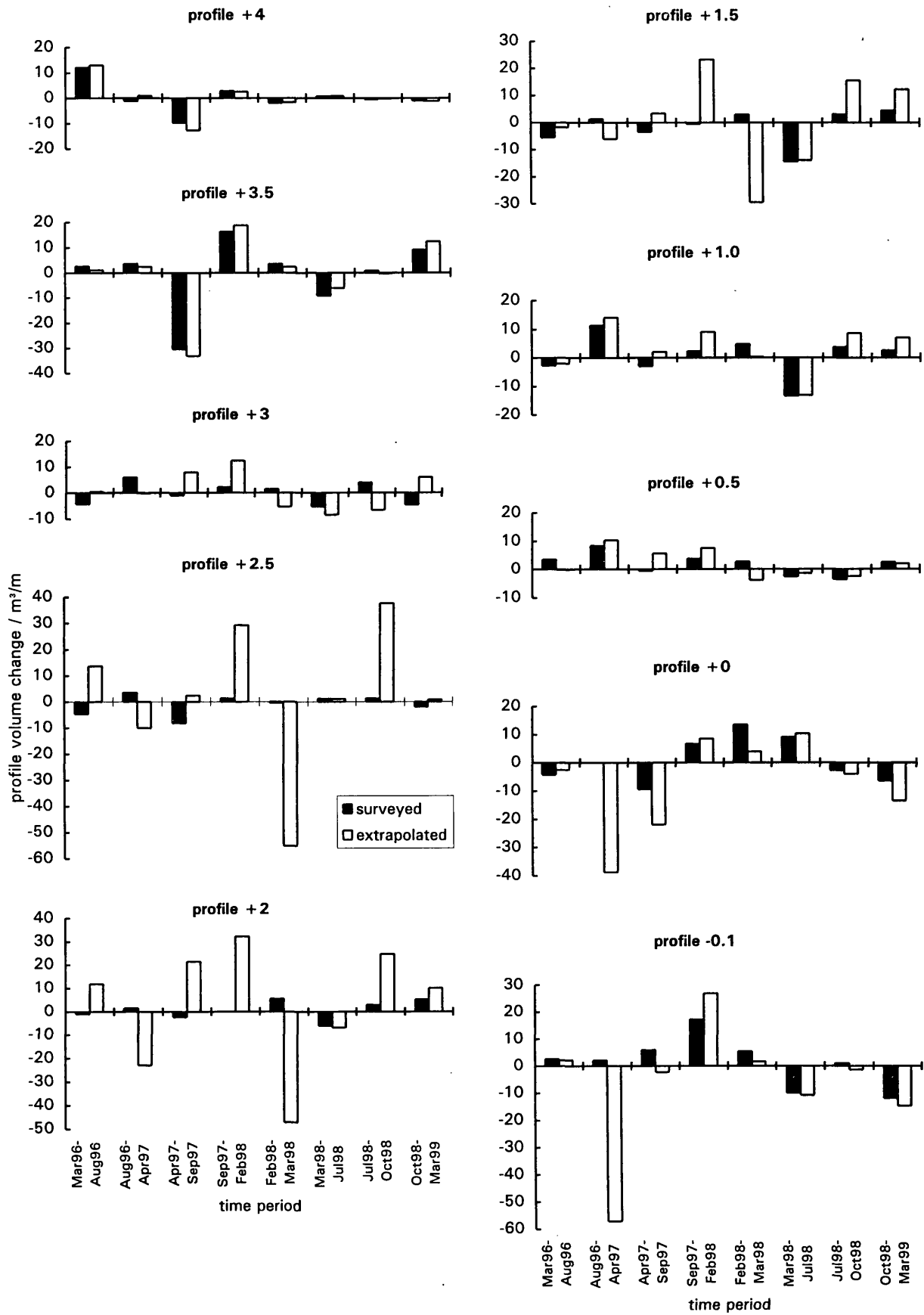


Figure 6.18 (continued): Volume changes at each profile in each survey period (in m^3/m)

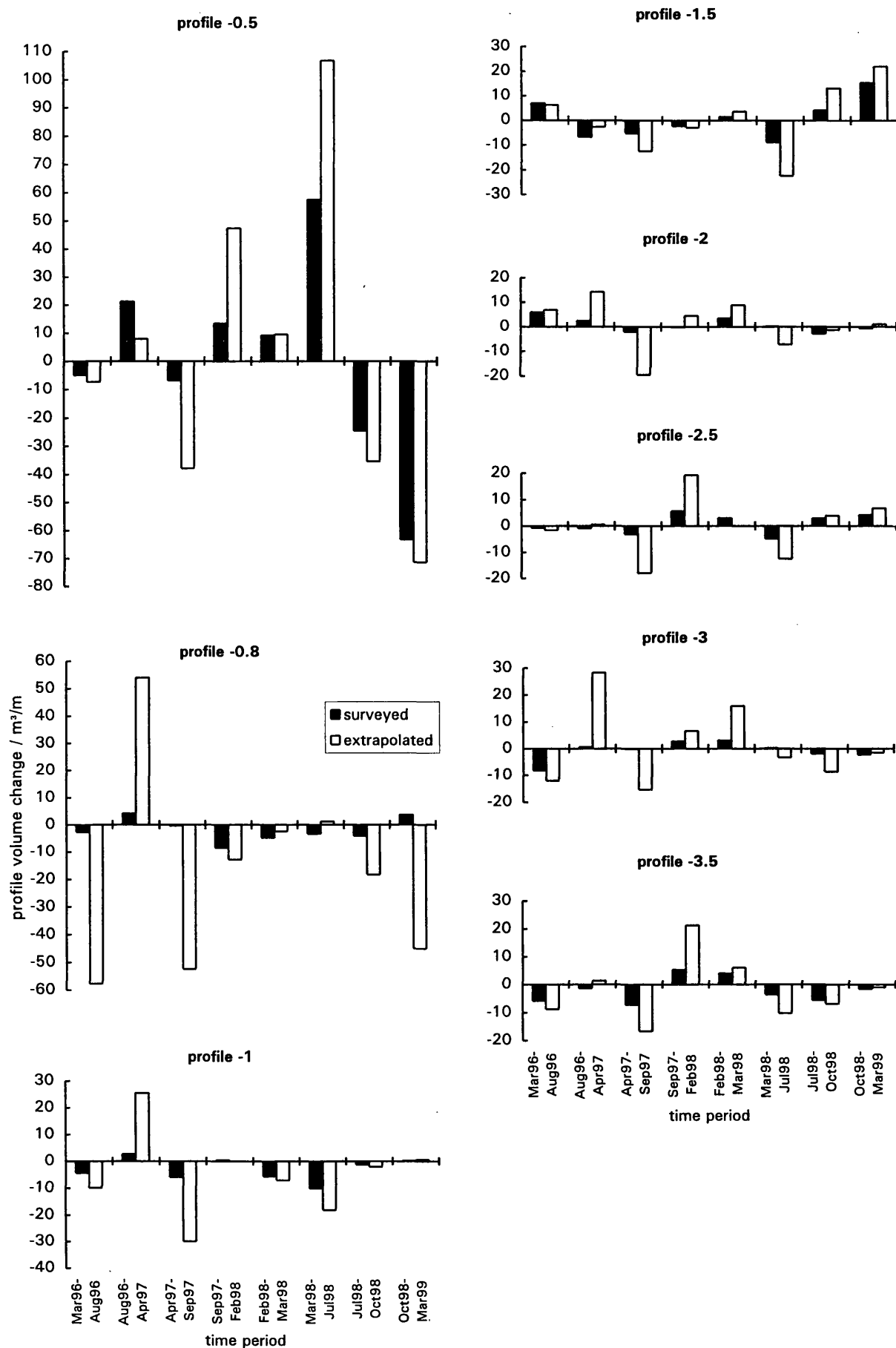


Figure 6.18 (continued): Volume changes at each profile in each survey period (in m^3/m)

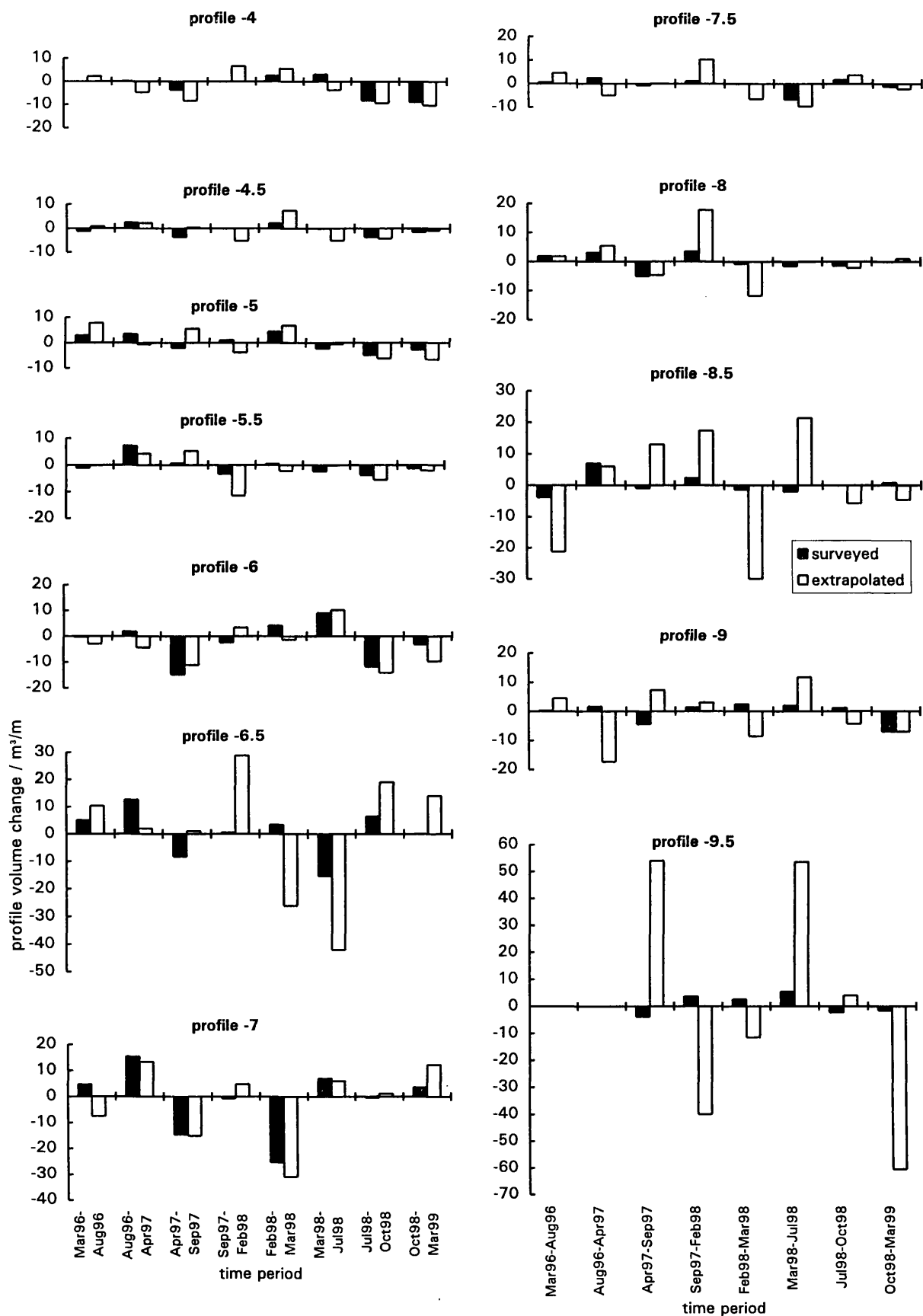
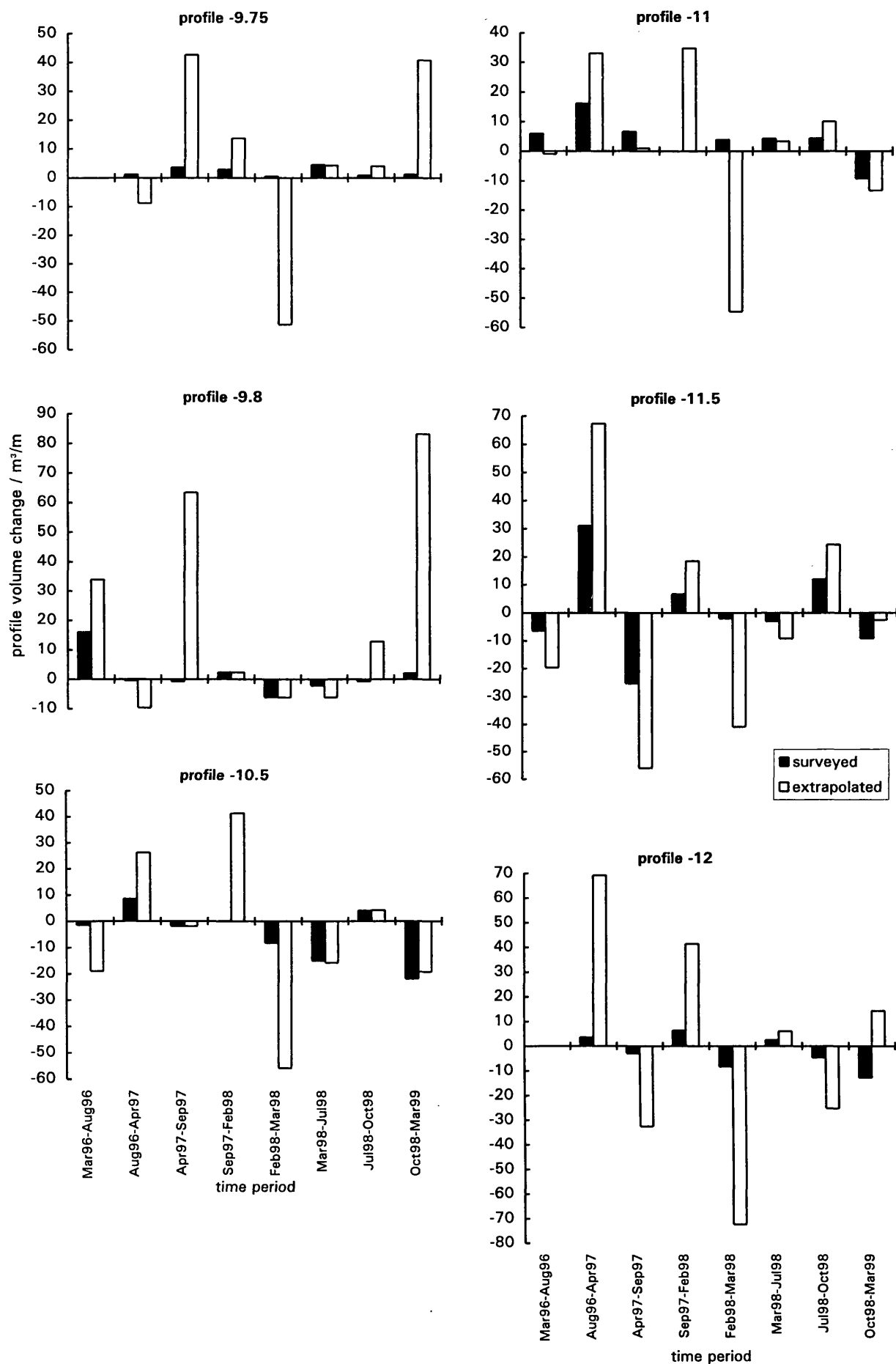


Figure 6.18 (continued): Volume changes at each profile in each survey period (in m^3/m)



The maximum actual volumetric gains and losses are $+58 \text{ m}^3/\text{m}$ and $-63 \text{ m}^3/\text{m}$, respectively, both of which were recorded at profile -0.5 (Figure 6.18). Using extrapolated profiles, the maximum volumetric changes are larger, at $+107 \text{ m}^3/\text{m}$ (profile -0.5) and $-72 \text{ m}^3/\text{m}$ (profile -12) (Figure 6.18).

These phases of erosion and accretion at each profile can be directly related to the changes in beach morphology documented in Section 6.3 and Appendix F. The profile volumetric changes presented herein (Figure 6.18) will be converted to cell volume changes for each ca. 500m wide cell in Section 6.6.2.3 and used to compute the coastal sediment budget. Given the potential errors involved when calculating volume changes from extrapolated profiles (approach 3), the sediment budget is compiled using only the changes quantified using surveyed profiles (approach 2).

6.6.2.2 Spatial patterns of beach profile volume changes

This section presents the recorded beach profile changes in a spatial context (Figure 6.19). In general, it can be seen that volumetric changes along the coast of Spey Bay are not spatially consistent. For example, the storm event on the 1st March 1998 appears to have created alternate zones of volumetric sediment gain and loss along the coastline (Figure 6.19). This spatially variable morphological response to the storm was noted in the discussion in Section 6.3. In addition, these zones of gain and loss do not appear to be static over time, with different parts of the beach experiencing erosion or accretion at different times. For example, compare the changes that occurred between February 1998 and March 1998 with those between July 1998 and October 1998. Zones of erosion during the first period became zones of accretion during the next. This may have implications concerning the operation of the coastal system.

Figure 6.19 also highlights some discrepancies between volumes calculated using the profiles as surveyed and those calculated by extrapolation down to a maximum limit. These differences become significant along certain stretches of the coastline (for example between profiles +1 and +3 and between -10 to -12). Extrapolation errors are likely to be greater on the sand beach (-10 to -12) and beaches with a sandy lower foreshore (+1 to +3) as the seaward limit between different surveys is more variable than on steeper gravel beaches. Again, the volume changes quantified using the surveyed profiles are considered more reliable.

Summing the short-term volumetric changes recorded over the three year period yields important information concerning the general trends of shoreline erosion or accretion along

Figure 6.19: Recorded beach volume changes (in m³/m) at each profile along the coast of Spey Bay for each survey interval.

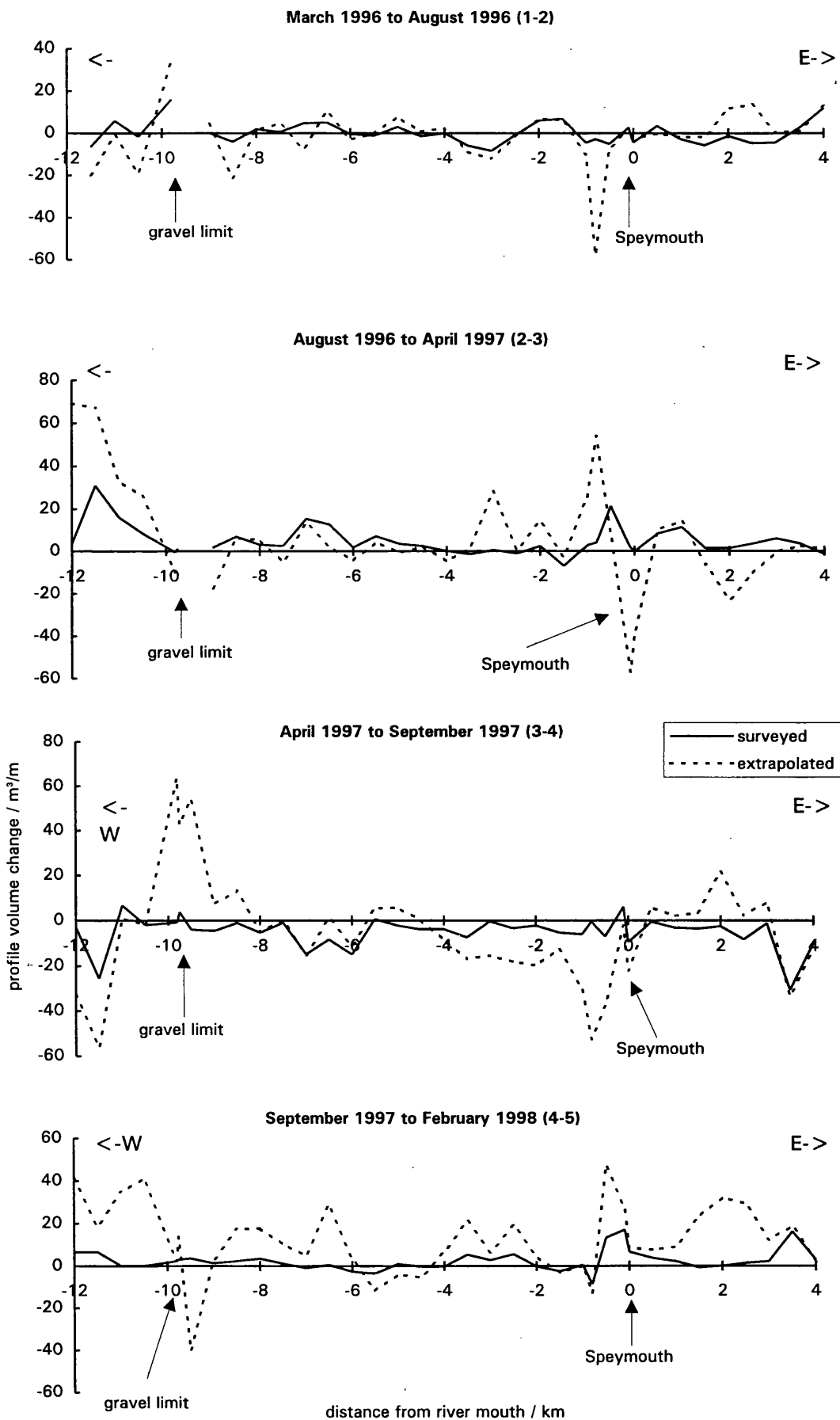
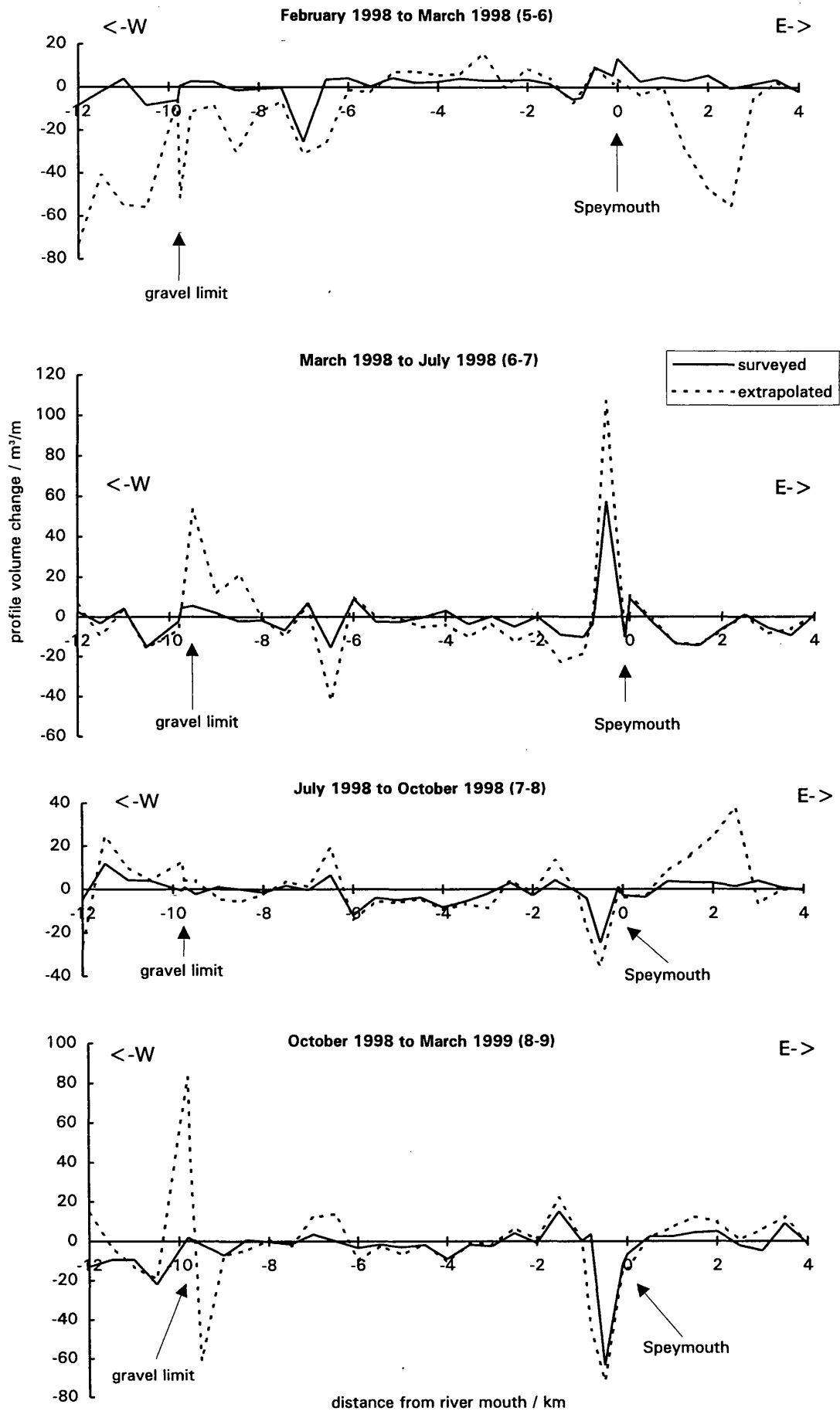


Figure 6.19 (continued): Recorded beach volume changes (in m^3/m) at each profile along the coast of Spey Bay for each survey interval.



the coast of Spey Bay (Figure 6.20). The beach exhibits alternating zones of erosion and accretion along the shoreline over the three years. For example, the beach was erosional between -0.8 and -1, accretionary along -1.5 to -2.5 and erosional along -3 to -5.5. Net volume changes are compared to the volumetric changes calculated by merely comparing the profiles as surveyed at the beginning (March 1996) and end (March 1999) of the study (Figure 6.20). The degree of similarity in the volumetric results obtained using these two methods indicates that the volume changes computed for each survey period (using the actual surveyed data without extrapolation) are robust and of reasonable quality. The beach profile analysis shows a similar trend of beach erosion and accretion as was identified from map analysis (Section 6.1). This is explored further in Chapter 7 when the medium-term (100 year) and short-term (3 year) budgets are presented in full.

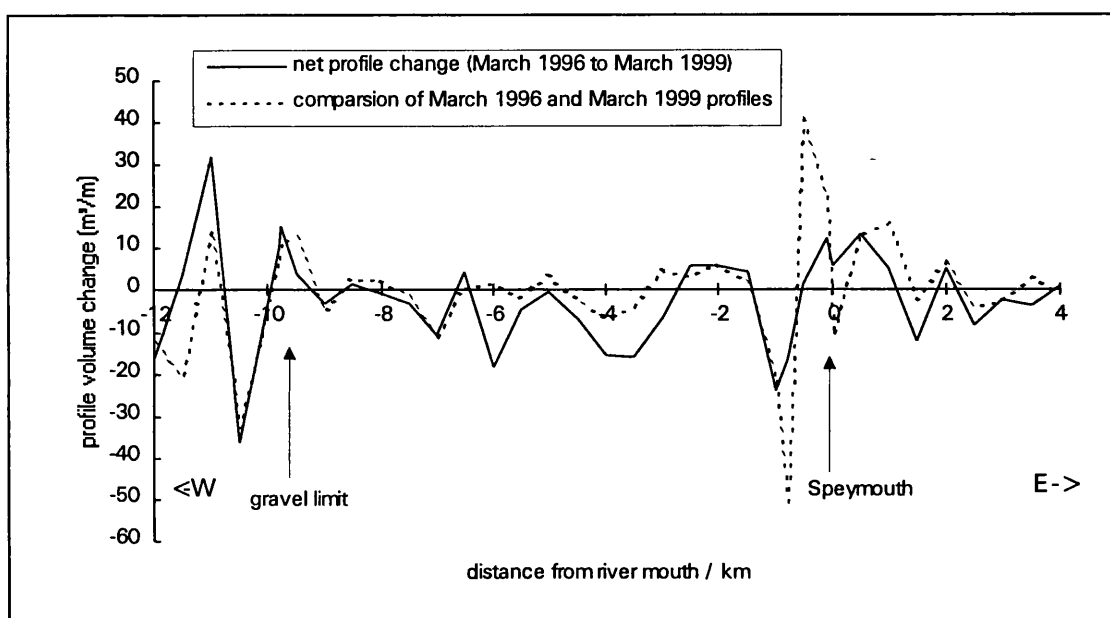


Figure 6.20: Net volumetric change (in m^3/m) at each profile along the coast of Spey Bay between March 1996 and March 1999. Volumetric change quantified using the two end profiles (March 1996 and March 1999) are shown for comparison. General trends of erosion and accretion can be identified.

6.6.2.3 Quantification of beach cell volume changes

Sediment volume changes for each ca. 500m cell of the Spey Bay coastline were calculated using the mean end areas approach (equation 2.11, Section 2.1.2) using the volumetric changes calculated using both the surveyed and extrapolated profiles (Appendix I). It is worth noting that, not only are the extrapolated cell volumes generally of much greater magnitude but, in some cases, they are also in the wrong direction, showing accretion instead of erosion (e.g. cell 2-2.5, April 1997 to September 1997). This raises a note of caution concerning the reliability of sediment budgets which are calculated based on

profiles extrapolated to a specified (and generally untested) closure depth. As a result, this section will discuss only the cell volume changes derived from surveyed profile data (Figure 6.21).

A robust check on the validity of using surveyed beach profiles to estimate volume changes within the cells they delimit is provided by the detailed repeat survey of the Kingston delta (at a survey point spacing of ca. 6m) between profiles -0.5 and -0.8 (Section 5.5.2). Using beach profile data the cell was estimated to have gained $10\,150\text{ m}^3$ of sediment between surveys 3 (April 1997) and survey 7 (July 1998) (Appendix I). This is in extremely close agreement to the best estimate of the actual volume change calculated from detailed surveys of the entire cell between May 1997 and May 1998 ($+10\,320\text{ m}^3$, Table 5.9). This internal consistency lends considerable authority to the validity of the technique used.

The beach showed a variable response to environmental conditions over the summer of 1996 (Figure 6.21a), with zones of the beach experiencing erosion and others accretion. In contrast, volumetric gain was experienced in almost all cells between August 1996 and April 1997 (Figure 6.21b) with up to $12\,000\text{ m}^3$ of sand accreting in the cell delimited by -11 and -11.5, mainly due to embryo dune accretion. This phase of accretion was reversed the following summer, when all but one cell experienced volumetric loss (Figure 6.21c). The following five survey periods show zones of the beach experiencing erosion alternating with zones of accretion (Figures 6.21d-h) and indicate the highly variable spatial and temporal response of the Spey Bay beaches to environmental conditions. Beach response to the storm of December 1997/early January 1998 (Section 3.5.1.5) varied spatially, with alternating zones of accretion and erosion at a spacing of ca. 4km along the beach (Figure 6.21d). Most of the beach experienced sediment gain following the storm of March 1998 (Figure 6.21e), with the exception of the cells between profiles -0.5 and -1.5, -6 and -9 and the sand beach at Lossiemouth (profile -9.8 to -12). This is not surprising, given that the morphological response to storms at Spey Bay is often one of crestal retreat and storm overwashing (Section 6.3), resulting in increased volumes on the upper beach.

The *net* cell volumetric changes over the three years (Figure 6.22) highlight that while relatively large changes in cell volume can occur between successive surveys (e.g. up to $-10\,800\text{ m}^3$ in the cell delimited by profiles -0.5 and -0.8) (Figure 6.21h) net changes over three years in each cell are relatively small (maximum $+9\,000\text{ m}^3$, cell -11 to -11.5). This is a result of alternating temporal and spatial phases of erosion and accretion. Some general trends can tentatively be suggested. Over the three years the eastern part of east Spey Bay ($+1$ to $+4$) has been generally erosional, while the beach closer to river mouth has

Figure 6.21: Volume changes (in m³) for each ca. 500m wide cell along Spey Bay between each survey period

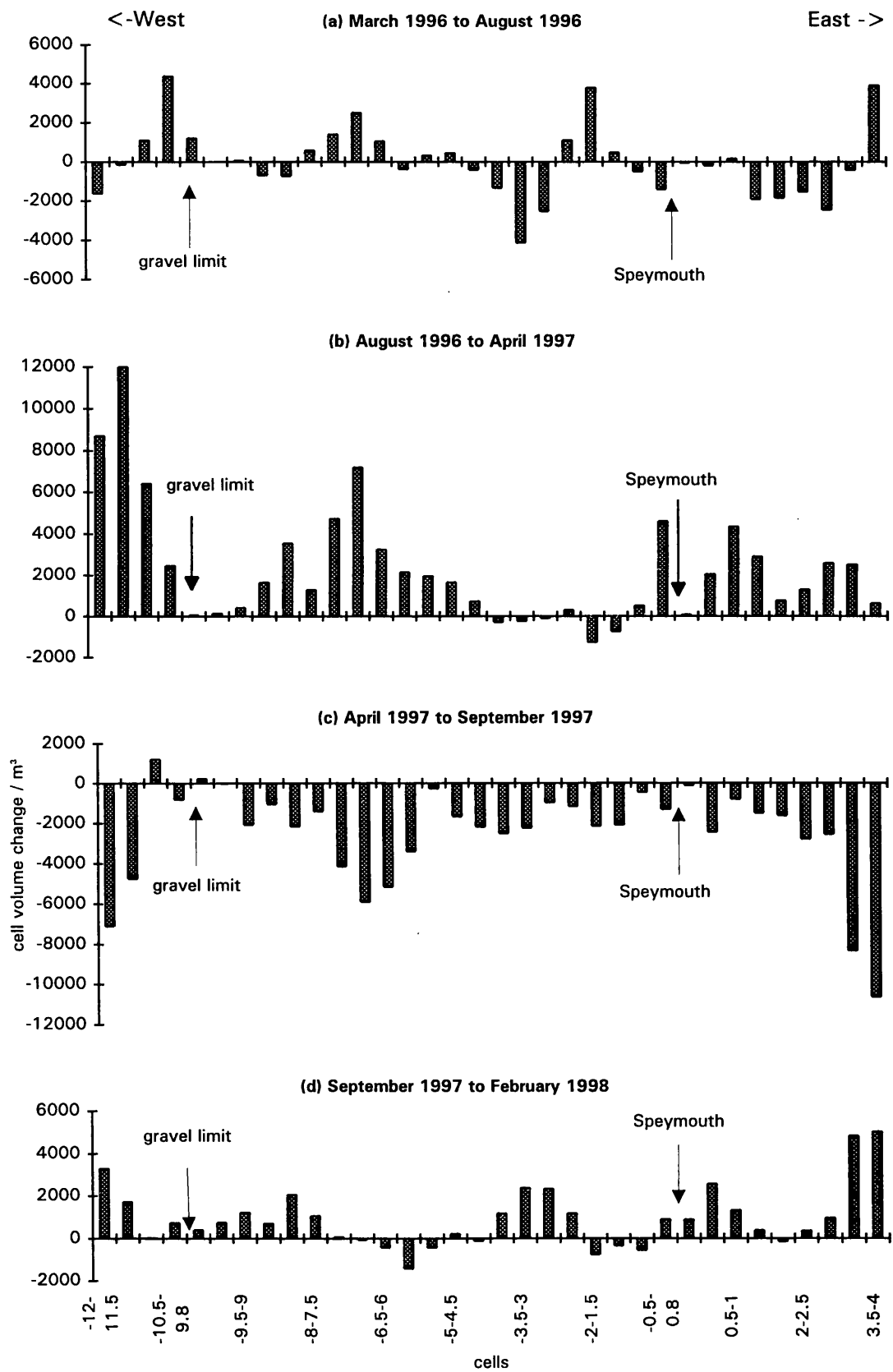
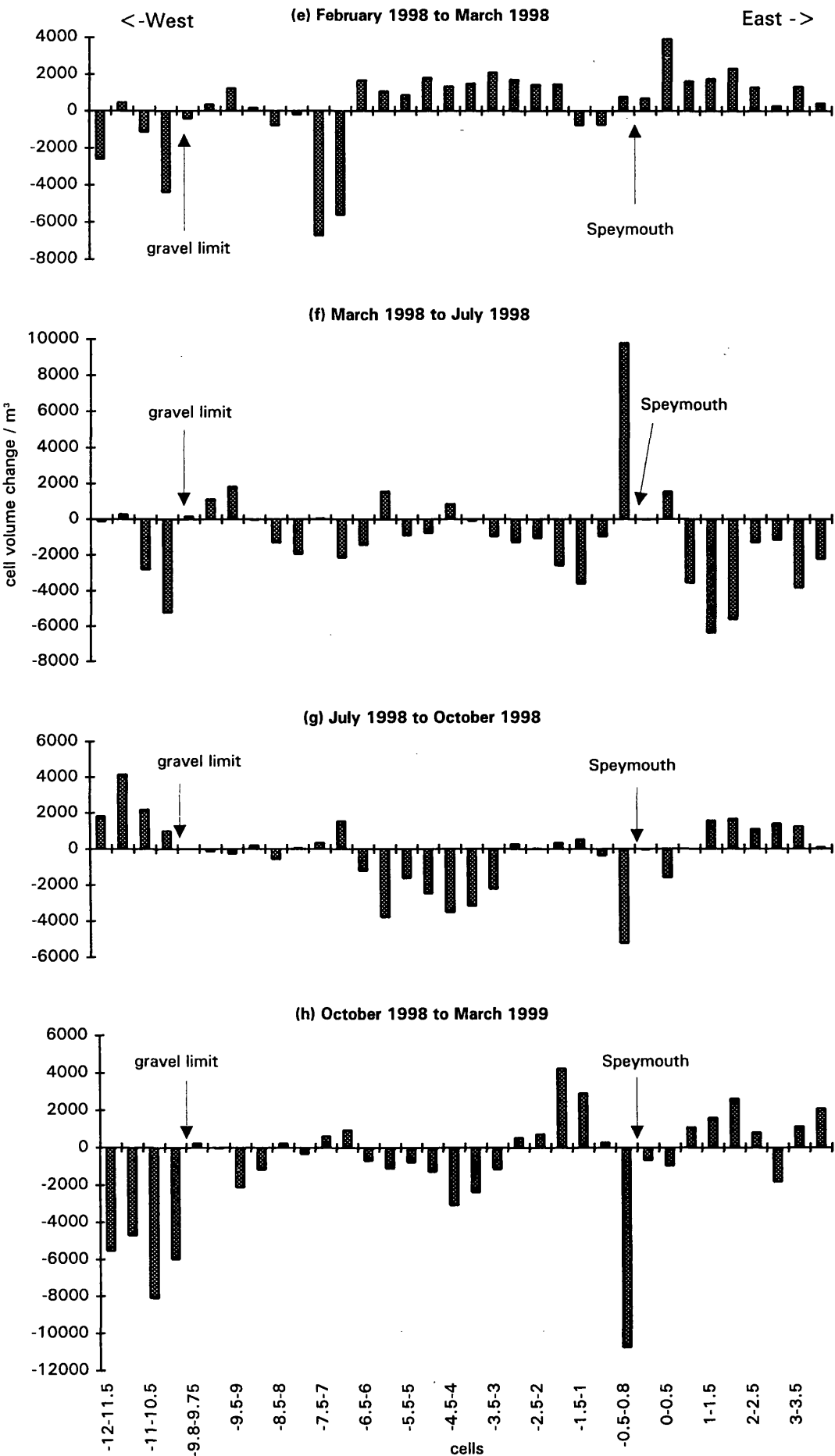


Figure 6.21 (continued): Volume changes (in m³) for each ca. 500m wide cell along Spey Bay between each survey period



experienced accretion, associated with deposition at the delta (Figure 6.22). Immediately west of the river mouth a zone of erosion stretches from the delta to profile -1.5. Net accretion over the three years occurred between profiles -1.5 and -2.5, with a major zone of net erosion extending westwards from profiles -3 to -8. The cells between profiles -8.5 and -9.8 experienced net accretion over the three years due to the westerly migration of the gravel beach (Section 6.1). The sand beach at Lossiemouth was generally erosional over the three year study period, with the exception of cell -11.5 to -11 (Figure 6.22) where the upper beach has accreted by wind blown sand. Summing all cell volume changes over the three years, the entire beach is estimated to have undergone a net loss of ca. 40 000m³ of sediment (ca. 13 000 m³a⁻¹) (Figure 6.22).

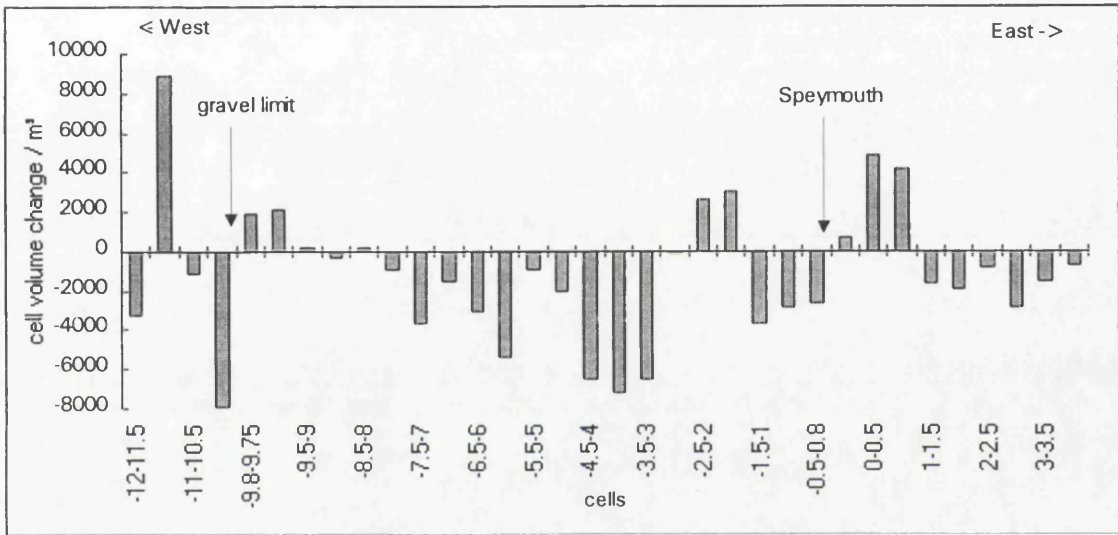


Figure 6.22: Net volumetric change for each cell along the coast of Spey Bay (in m³). The entire beach is estimated to have undergone a net loss of ca. 40 000m³ of sediment between March 1996 and March 1999.

6.6.2.4 Error analysis

The calculated beach cell volume changes (Figure 6.21 and Appendix I) are estimates of the actual storage change occurring in each cell, and, as such, have associated uncertainties. The uncertainty is a combination of surveying errors and bias introduced by the assumption that the area change at a profile is representative over the distance between it and the half distance to each adjacent profile (i.e. errors associated with beach profile spacing, Section 4.6.2).

Surveying Errors

Beach surveys used levelling (Section 4.3.1), the accuracy of which is of the order of $\pm 0.05\text{m}$ (Gable and Wanetick 1984). This error (δ) in m can be converted to an area error (δA) (in m²) as it is related to the area change at each beach profile (A) by:

$$\delta A (m^2) = (\sqrt{A} + \delta)^2 - A = 2\delta\sqrt{A} + \delta^2 \quad (6.1)$$

Area change errors (m^2) at each profile are converted into cell storage change errors, δV (m^3), using:

$$\delta V = \frac{\delta A_i + \delta A_{(i+1)}}{2} L_{(i,i+1)} \quad (6.2)$$

where δA_i is the area change error at profile i , δA_{i+1} is the area change error at an adjacent profile and $L_{(i,i+1)}$ is the distance between two profiles (i.e. cell width).

Uncertainties due to beach profile spacing

Detailed analysis of beach profile spacing (Section 4.6.2) highlighted the sensitivity of volume change calculations to profile spacing. The mean percentage error in volume change calculations associated with a profile spacing of 500m is 118% (Section 4.6.2.8, Figure 4.18). This error is calculated from analysis of actual changes recorded at profile -2.0. However, in this test the actual change in area at each profile between surveys was very small (mean change was ca. $7 m^3/m$, Table 4.3) producing the large relative volume error.

Nevertheless, the mean absolute area change of all profile change data is of a similar order of magnitude (ca. $5 m^3/m$) and so a volumetric error of $\pm 118\%$ is applied to all cell volume changes. It should be emphasised that this is likely to over-estimate the real errors, especially as the scale of the change increases (Section 4.6.2.8). For large volume changes associated with major profile shifts the percentage error will be much smaller.

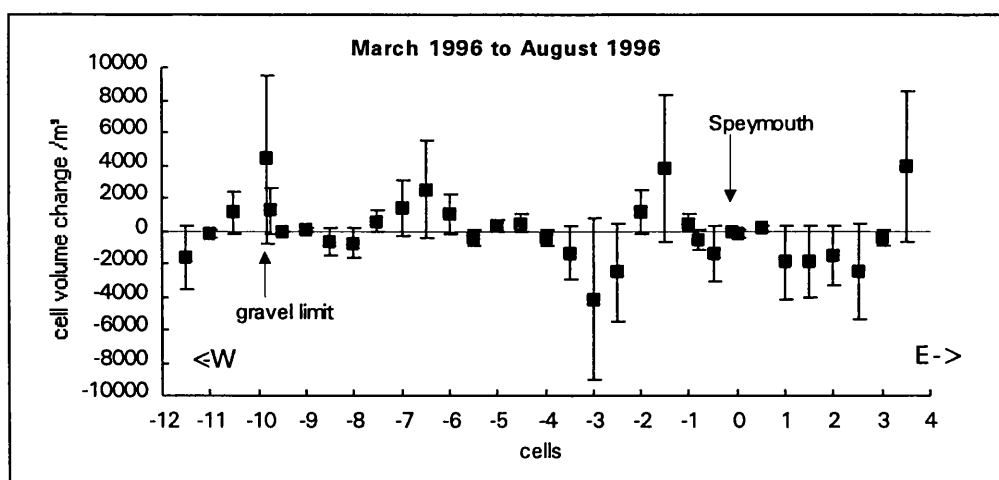


Figure 6.23: Cell volume changes (in m^3) for each cell between survey 1 (March 1996) and survey 2 (August 1996) (as presented in Figure 6.21a) with maximum error bars. Errors are a combination of surveying error and bias associated with beach profile spacing.

Surveying and spacing errors were calculated and combined to quantify the RMS error (in m^3) of volume change estimates for each cell at each time period (Appendix J). Errors in the estimation of cell volume changes can be significant (Figure 6.23) and are always larger than the estimated volume change. These errors will be accounted for when the contemporary Spey Bay sediment budget is compiled (Chapter 7). It is stressed that these are maximum errors and, in reality are likely to be much smaller, particularly for larger changes.

6.7 Summary

Beach volume change at Spey Bay has been estimated using successive beach profile surveys and suggests a net sediment loss of ca. $13\,000\text{ m}^3\text{a}^{-1}$ between March 1996 and March 1999. Results show that the beach exhibits alternate zones of accretion and erosion, which vary spatially and temporally along the coast. Within this variability, some general trends in beach erosion and accretion are identified which corroborate medium-term trends identified from map analysis (1870-1970).

7. DISCUSSION

This chapter presents a short-term (3 year) sediment budget for the Spey Bay system and evaluates the methodology used. The budget is compiled by combining fluvial and deltaic sediment transfers (Chapter 5) with coastal sediment transfers (Chapter 6) over the period from 1996 to 1999. Longshore gravel transport rates, estimated using the principle of sediment continuity, are presented and evaluated. Medium-term (100 year) and Holocene (10 000 year) budgets are also presented and discussed. The temporal and spatial variability in fluvial, deltaic and coastal sediment transfers (Chapters 5 and 6) are used to evaluate the nature of sediment exchanges and investigate the processes and mechanisms by which the Spey Bay system operates. A conceptual model of gravel sediment transfers from rivers to beaches is proposed and areas for potential future research are identified.

7.1 Spey Bay sediment budgets

Sediment budgets can be constructed for several time-scales (Section 2.1); appropriate scales depend on both process considerations and the quality of evidence. Long-term (Holocene) budgets are affected by sea-level changes and variations in sediment supply and are generally based on geomorphological reconstructions (e.g. Comber 1993). Medium-term budgets, constructed using evidence for ca. 100 years, identify the general trends of system evolution and are probably the best scale to use when trying to make future predictions and management decisions. Short-term budgets give the details and mechanisms for the two longer scales and are required to provide accurate predictions of future changes and assess the nature of system operation. The desired temporal scale has implications for how sediment budgets are constructed (e.g. level of detail, quality of evidence and accuracy of estimates).

This study has concentrated on the medium and short-term scales. A medium-term budget compiled from map analysis and river modelling is used to gain an insight into the trend of system evolution at Spey Bay. This is presented first to set the short-term changes recorded between 1996 and 1999 into context. The contemporary (short-term) budget follows in Section 7.1.2 and is used to assess the details, mechanisms and nature of system operation. Estimation of longer-term budget changes over the Holocene and how they compare with the shorter trends are discussed briefly in Section 7.2.

7.1.1 Medium-term (historical) budget (1870-1970)

The average annual gains and losses of sediment in the coastal system are estimated from map and volumetric analysis presented in Section 6.2. Although Spey Bay as a whole experienced a minimum total loss of $1\,676\,000\text{ m}^3$ between 1870 and 1970, representing an annual loss of ca. $17\,000\text{ m}^3\text{a}^{-1}$, the section between profiles -4.5 and -8 displayed volumetric gains of ca. $6\,000\text{ m}^3\text{a}^{-1}$ (Table 6.2). River input is estimated from river modelling studies (Dobbie & Partners 1990) which suggest that the River Spey contributes an average of $8\,000\text{ m}^3\text{a}^{-1}$ of sediment to the coastal system, although this is variable depending on flow conditions in any given year (Section 5.4.1). Delta storage is not considered at this scale. All volumes are expressed as mineral volumes (i.e. corrected for porosity).

The porosity of coarse-clastic beaches is difficult to measure, as most contain varying mixtures of gravel and sand, alongshore, cross-beach and at depth (Carter and Orford 1984) causing spatial variation in porosity. No values for porosity of coarse-clastic beaches are given in the literature, but since the Spey beaches are sourced from the River Spey (Section 3.3) it is assumed herein that the beach and delta sediment have a similar porosity to the river gravels ($\rho = 0.3 \pm 0.05$, Section 5.3.3). Published values of porosity for sand beaches vary from 0.39 (Tickell and Hiatt 1938) to 0.49 (Pryor 1973). A value of 0.45 ± 0.05 is used for the sand beach at Spey Bay, which is the mean of all published values for beach sand with a similar grain size and sorting (Atkins and McBride 1992, p340).

Since sand and gravel are transported by different mechanisms and ultimately have different fates within Spey Bay, they are treated separately in the sediment budget calculations. Bulk sediment samples indicate that the input source sediment from the River Spey contains 80% gravel and 20% sand (Section 5.3.2), whereas sediment transfers on the coarse-clastic beaches contain 95% gravel and 5% sand (Section 6.5.2). It is assumed that most of the fluvial sand is transferred via the nearshore and offshore zones and is not involved in beach face sediment transfers. The volume changes in Table 6.2 and the annual fluvial input are converted to mineral volumes of gravel and sand and used to construct the medium-term budget for the Spey Bay system (Table 7.1 and Figure 7.1).

Gravel inputs into Spey Bay come from the river and from recycling of beach gravels by erosion (Table 7.1 and Figure 7.1). A total of $14\,300\text{ m}^3\text{a}^{-1}$ of gravel is input into the beach system. The output or sink volumes relate to the build-up of gravels at Boar's Head Rock which, using map evidence, amounts to $4\,000\text{ m}^3\text{a}^{-1}$ above LWM. There is thus an apparent

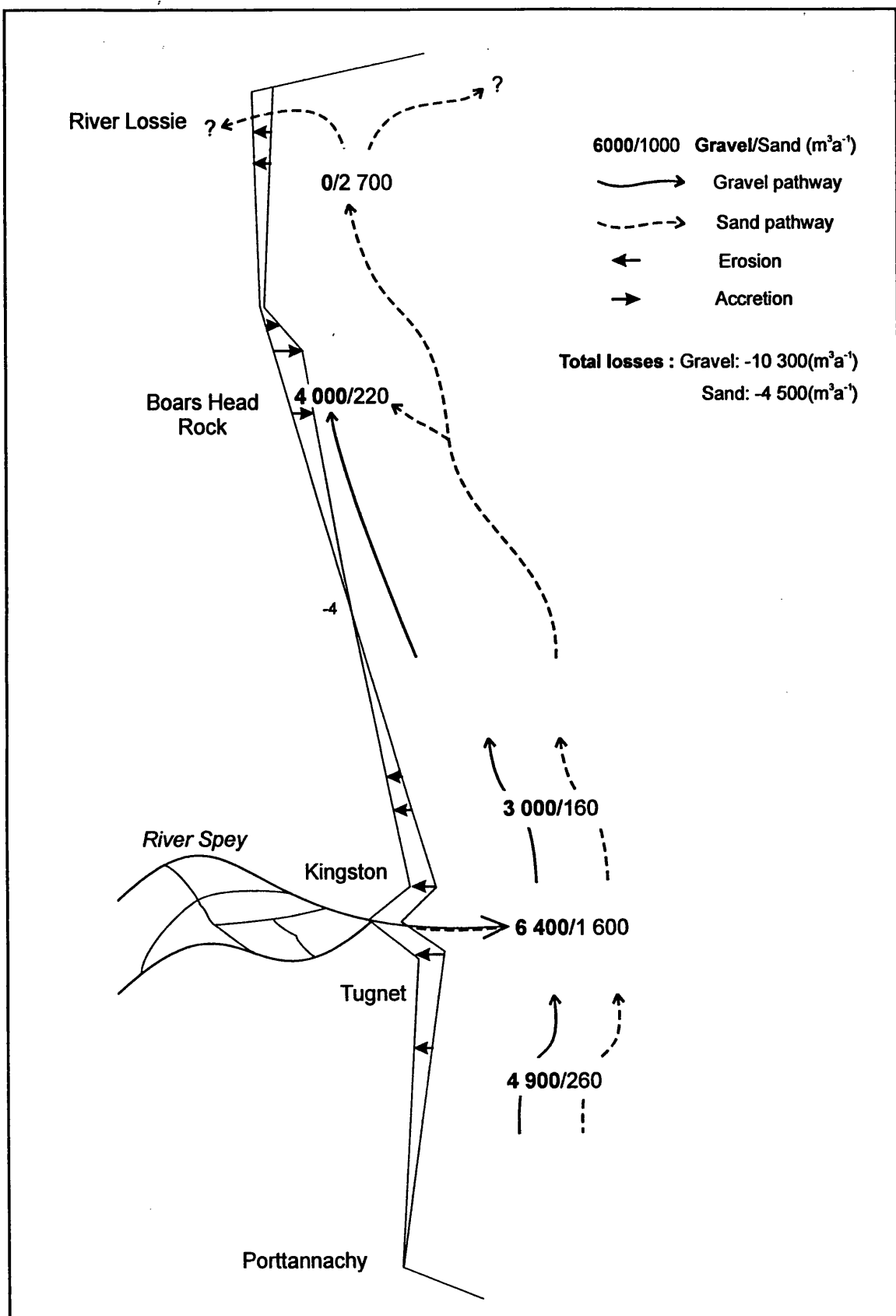


Figure 7.1: Diagrammatic representation of the Spey Bay medium-term sediment budget (1870 - 1970) compiled using data from river modelling studies (Dobbie & Partners 1990) and map analysis (Section 6.2).

net loss of $10\,300\text{ m}^3\text{ a}^{-1}$ of gravel from the Spey system. Whilst this may involve errors in the delivery of Spey gravels, it is more likely that much of the error lies in the amount of sediment gained and lost below L.W.M, largely unaccounted for in this analysis.

	Gravel		Sand	
	Source	Sink	Source	Sink
Coastal Erosion (East Spey Bay)	+ 4 900		+ 260	
River Spey supply	+ 6 400		+ 1 600	
Coastal Erosion (West Spey Bay to profile -4.0)	+ 3 000		+ 160	
Accretion near Boar's Head Rock (from profiles -4.0 to -8.5)		- 4 000		- 220
Erosion (from profiles -8.5 to -12)			+ 2 700	
TOTAL	+ 14 300	- 4 000	+ 4 720	- 220
BUDGET - NET DEFICIT	10 300		4 500	

Table 7.1: Medium term gravel and sand sediment budget for Spey Bay in $\text{m}^3\text{ a}^{-1}$ (mineral volumes) for the period 1870 to 1970.

There is also a net loss of sand from the Lossiemouth end of the beach to the west of profile -8.5, which averages $2\,700\text{ m}^3\text{ a}^{-1}$ between 1870 and 1970 (Table 7.1 and Figure 7.1). This is augmented by the sand fraction derived from erosion of the updrift coast and from the Spey supply to produce a net loss of $4\,500\text{ m}^3\text{ a}^{-1}$. The final destination of this sand is likely to be both sand dune accretion behind the beach and infill of the Lossie saltings, together with an unknown loss to the offshore zone. It is also possible that there is sediment leakage round the headland of Branderburgh, although H.R. Wallingford (1995) suggest that this may be unlikely.

The budget shows the general trends of beach erosion and accretion at Spey Bay. In summary, the map data show that the contribution of material from the alongshore, offshore and river combined is about $19\,000\text{ m}^3\text{ a}^{-1}$ and the net loss is about $14\,800\text{ m}^3\text{ a}^{-1}$. Although the coastline near Boar's Head Rock shows some accretion, the rest of the coast is erosional. These estimates will be compared to the contemporary budget compiled using the morphological data presented in Chapters 5 and 6.

7.1.2 Short-term (contemporary) budget (1996-1999)

7.1.2.1 Net changes (1996-1999)

The contemporary budget for Spey Bay (Table 7.2 and Figure 7.2) is quantified using fluvial sediment transfers (Section 5.4.2 and Appendix B), deltaic transfers and storage (Section 5.5) and the net beach cell storage changes recorded from March 1996 to March 1999 (Figure 6.22 and Appendix I). Annual rates of sediment transport from the River Spey to the coast over this period vary from ca. $20\,000\text{ m}^3\text{ a}^{-1}$ (December 1995 to December 1997) to ca. $6\,000\text{ m}^3\text{ a}^{-1}$ (December 1997 to April 1999) (Section 5.4.2). The mean annual

transfer to the coast between March 1996 and March 1999 is calculated to be ca. 13 750 m³ a⁻¹ (mineral volume).

	GRAVEL		SAND	
	source	sink	source	sink
Erosion (east) - profiles +4 to +1	+2 000		+100	
Accretion (east) - profiles +1 to -0.1		-2 100		-100
River Spey supply	+11 000		+2 800	
Accretion at the delta		-2 900		-150
Erosion (west) profiles -0.5 to -1.5	+2 000		+100	
Accretion (west) profiles -1.5 -2.5		-1 200		-100
Erosion (west) profiles -2.5 to -8	+8 400		+400	
Accretion (west) profiles -8 to -9.8		-900		-50
Erosion (Lossiemouth sand beach profiles -9.8 to -12)			+600	
TOTAL	+23 400	-7 100	+4 000	-400
BUDGET - NET DEFICIT	16 300		3 600	

Table 7.2: Contemporary gravel and sand sediment budget for Spey Bay in m³ a⁻¹ (mineral volumes) for the period March 1996 to March 1999. See text for details.

The contemporary budget shows an apparent net loss of sediment from the system, amounting to ca. 16 300 m³ a⁻¹ and 3 600 m³ a⁻¹ of gravel and sand, respectively. Estimates from the medium-term budget give very similar figures (gravel and sand losses of ca. 10 300 m³ a⁻¹ of 4 500 m³ a⁻¹, respectively, Table 7.1). The consistency between contemporary and medium term budgets suggests that analysis of repeat beach profiles is a robust and valid technique for quantifying changes in sediment storage on the upper beach. However, as both budgets are in deficit, it is possible that on/offshore transfers of sediment are significant at both time-scales and that the apparent deficit represents errors in the amount of sediment gained and lost below LWM. The implications of this are investigated in Section 7.1.2.3.

The contemporary budget shows a total of ca. 23 400 m³ a⁻¹ of gravel input to the Spey system from the river and erosion of the beach gravels (Table 7.2). Most of the coastline is erosional, except for three short sections which have undergone short-term net accretion (Figure 7.2). The short zones of accretion at Tugnet (just east of the river mouth) and between profiles -1.5 and -2.5 (west of the river mouth) are inconsistent with the general medium-term pattern of erosion and accretion at Spey Bay (Figure 7.1). However, the proximity of these zones to the river mouth suggests a causal link. This accretion, together with beach accretion at the westerly migrating gravel front and the delta accounts for ca. 7 100 m³ a⁻¹ of the total input, leaving an average annual deficit of ca. -16 300 m³ of gravel.

The contemporary budget also shows that erosion of the sand beach at Lossiemouth, identified earlier from map analysis (Sections 6.1 and 7.1) is reflected in the short-term survey data, although smaller volumes are involved. The ca. 600 m³ a⁻¹ of sand input from

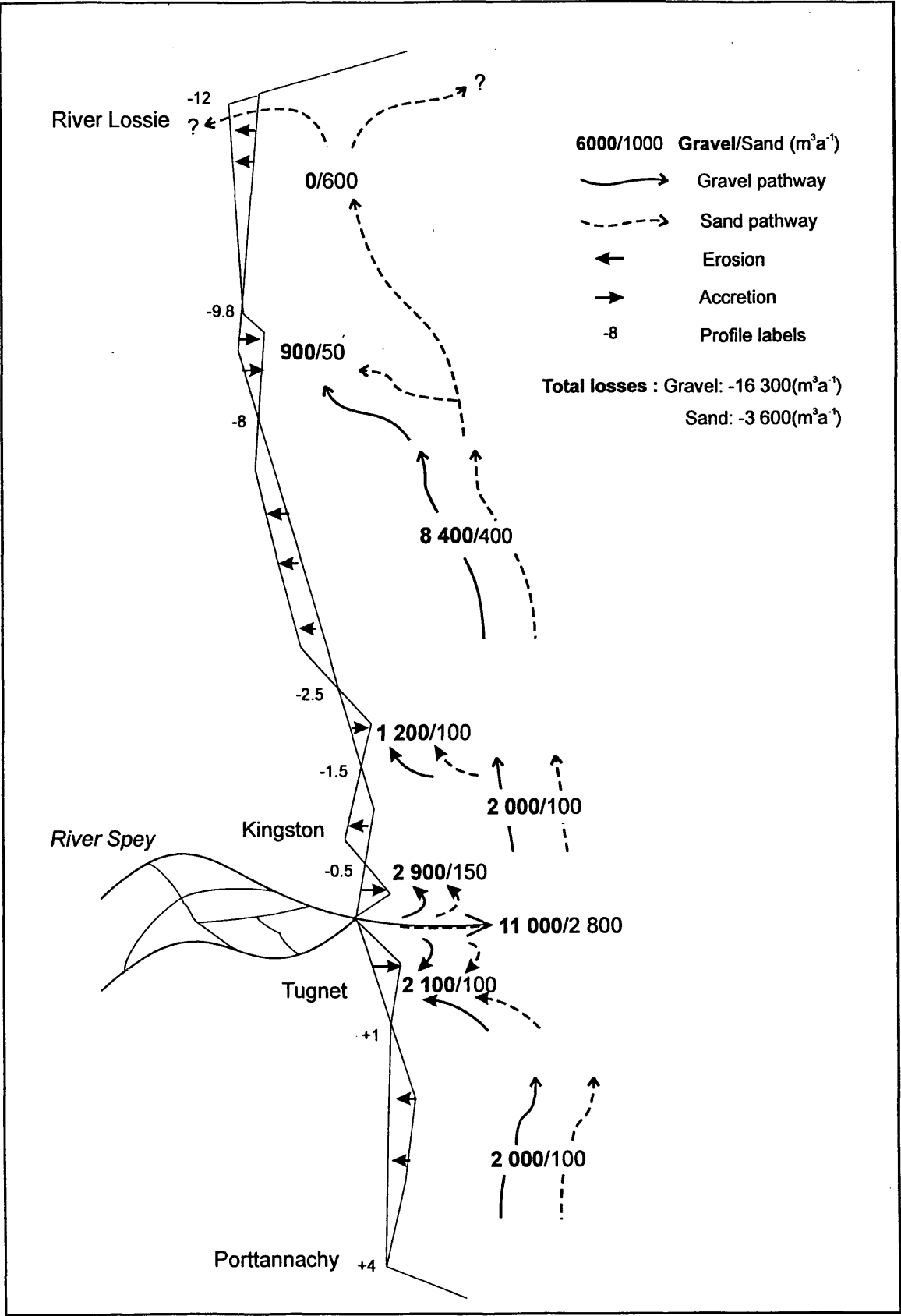


Figure 7.2: Diagrammatic representation of the Spey Bay short-term sediment budget (1996-1999) compiled using results from repeat morphological surveys of the river, delta and beach.

this stretch of the beach is augmented by the sand fraction from the river and beach erosion to give a total sand input to Spey Bay of ca. 4 000 m³a⁻¹. Recorded beach accretion accounts for less than 10% of the input, leaving a net annual deficit of sand of ca. 3 600 m³. This may be accounted for by unknown losses offshore, sand infill of the dunes and Lossie saltings or leakage around the headland.

7.1.2.2 Calculation of the rate of longshore gravel transport (1996-1999)

Applying the principle of sediment continuity, preliminary estimates of the volumetric rate of longshore gravel transport can be calculated using:

$$Q_o = Q_i - (1 - \rho) \delta S / \delta t \pm Q_r \quad (2.10, \text{chapter 2})$$

where Q_i and Q_o are the volumetric transport rates into and out of, respectively, each cell per unit time (δ), δS is the change in cell sediment storage and ρ is the sediment porosity. Q_r is a term accounting for other sources or sinks of beach sediment to a given cell (e.g. river input and on/offshore sediment exchanges). The applicability of this equation to coastal environments is discussed in Chapter 2.

Several assumptions are made when applying equation 2.10 to Spey Bay. Firstly, it is assumed that all transfers of sediment between beach cells are to the west (i.e. longshore drift is uni-directional). As the dominant waves arrive from the north-east sector due to the incidence of swell waves and greater fetch lengths from this direction (Section 3.5.1) it can be assumed that most sediment transfers are to the west. Storms from the west are generally short-lived and of lower magnitude (due to decreased fetch lengths) but may reverse the drift direction for short periods. Wave refraction analysis (Dobbie & Partners 1990) and field evidence confirm a strong westerly longshore drift along the Spey Bay coast (Section 3.5.1.4). Secondly, and more easily defined is an updrift boundary condition of $Q_i = 0$ at profile +4. This marks the eastern limit of the gravel beach and is bounded by the harbour and seawall of Porttannachy to the east. Any gravel input from the east is thus negligible. Thirdly, on/offshore exchanges of gravel are not considered in this analysis and Q_r is made up of river input and delta exchanges only. It seems reasonable that such exchanges may be negligible due to the very distinct gravel-sand transition observed at most profiles on the lower foreshore at ca. LWST (Section 6.5.1).

Sediment transfers between cells are calculated downdrift to the mouth of the Spey using equation 2.10. The input of sediment from the river is calculated using the transport rates quantified in Chapter 5 (19 500 ± 3 700 m³a⁻¹ for the period December 1995 to December 1997 and 6 000 ± 4 000 m³a⁻¹ for December 1997 to April 1999). Sediment accretion at the

subaerial delta is assumed to continue at the same rate as that determined for the period May 1997 to May 1998 ($4\,500 \pm 200 \text{ m}^3 \text{ a}^{-1}$). These rates are used to calculate the total volume of sediment input and output from the coastal system during each beach survey period. The assumption that rates remain constant is an oversimplification, but represents the best available estimate.

Gravel budgets for each period are presented in Figure 7.3. All apparent gravel transport rates increase at the river mouth as fluvial sediment enters the coastal system, with the exception of the period March 1998 to July 1998 when accretion in cell -0.5 to -0.8 (Figure 6.21f) causes a decrease (Figure 7.3). The beach west of profile -9.8 is composed entirely of sand (Section 6.5), so gravel transport past this point equals zero on all occasions. However, longshore gravel rates calculated using equation 2.10 with the above assumptions give a mean rate of $17\,000 \text{ m}^3 \text{ a}^{-1}$ out of cell -9.8 (ranging between $126\,000 \text{ m}^3 \text{ a}^{-1}$ and $-72\,000 \text{ m}^3 \text{ a}^{-1}$) (Figure 7.4). The implications of this will be discussed in the following section.

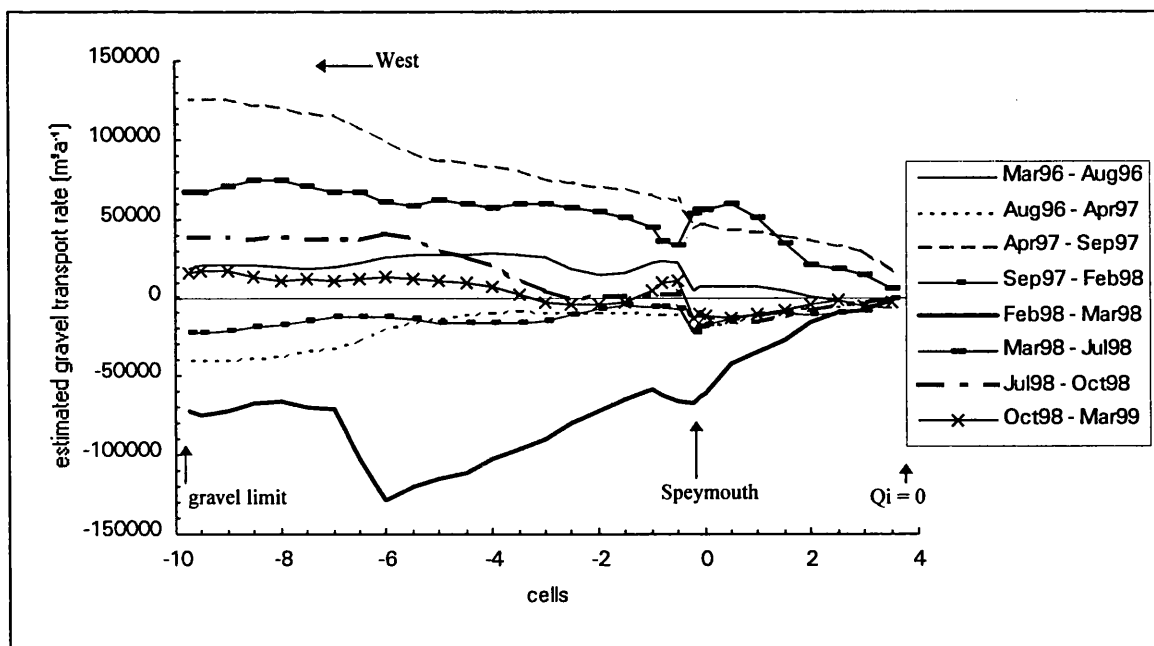


Figure 7.3: Longshore rates of gravel transport (in $\text{m}^3 \text{ a}^{-1}$) for Spey Bay, calculated using the equation of sediment continuity and assuming a zero boundary condition at profile +4 and a westerly longshore drift.

Apparent negative transport occurs when volumes of accretion (sinks) exceed volumes of erosion and river input (sources). This occurs on three occasions during the study (Figure 7.4) corresponding with periods of mean upper beach accretion (Figures 6.21b, d and e). Conversely, high apparent transport rates occur when source sediment (i.e. beach erosion and river input) greatly exceeds sink sediment (i.e. beach accretion). Phases of apparent

negative and positive transport appear to alternate through time (Figure 7.4) suggesting that there may be unknown on/offshore sediment transfers occurring. This budget, in common with many coastal sediment budgets, does not quantify this exchange and so assumes that these exchanges are equal to those required to balance the budget.

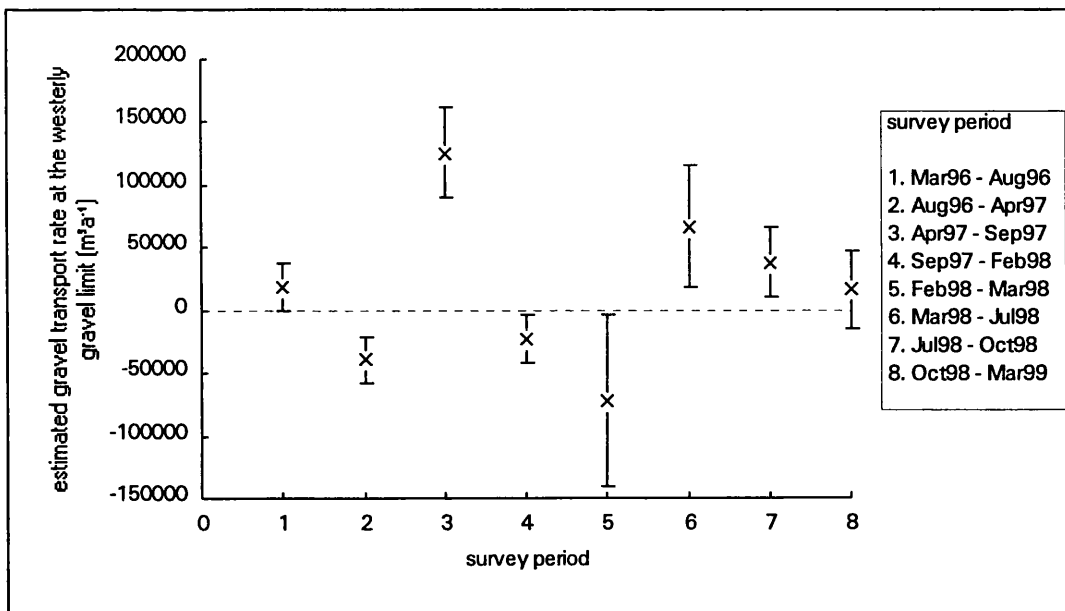


Figure 7.4: Estimated longshore rate of gravel transport at profile -9.8 using equation 7.1. Field evidence suggests this gravel transport past profile -9.8 is zero (see text).

7.1.2.3 Error analysis and limitations

The uncertainties in coastal sediment budgets must be assessed (Rosati and Kraus 1999). Uncertainties in fluvial transport rates are quantified in Section 5.4.4 and uncertainties in the estimates of delta and beach cell storage volume changes are given in Sections 5.5.2 and 6.6.2.4, respectively. Additional errors in storage change estimates are due to the porosity assumption and the grain size multiplier (i.e. the assumption that fluvial and beach sediment transfers contain 80% and 95% gravel, respectively). The porosity estimate was assigned an error range of ± 0.05 (cf. Martin and Church 1995) and the grain size multiplier was assumed to be within $\pm 3\%$ (Section 5.3.2).

Errors propagate downdrift, as the transport rate out of each cell (Q_o) is the transport rate into the downdrift cell, Q_i (equation 2.10). Cumulative errors in the transport rates are calculated in a similar way to those in the fluvial system (Section 5.4.4) using the equation (following Martin and Church 1995):

$$E_n = \sqrt{(\delta Q_{i_n})^2 + (\delta \rho_n)^2 + (\delta \Delta S_n)^2 + (\delta g_n)^2} \quad (7.1)$$

where δQi is the uncertainty in the transport rate into cell n , $\delta \rho$ is the uncertainty of the porosity, $\delta \Delta S$ is the uncertainty of the storage change estimate and δg is the uncertainty in the grain size multiplier for cell n . Downdrift of the river mouth, the uncertainty associated with the transport of fluvial sediment into the coastal system is added cumulatively.

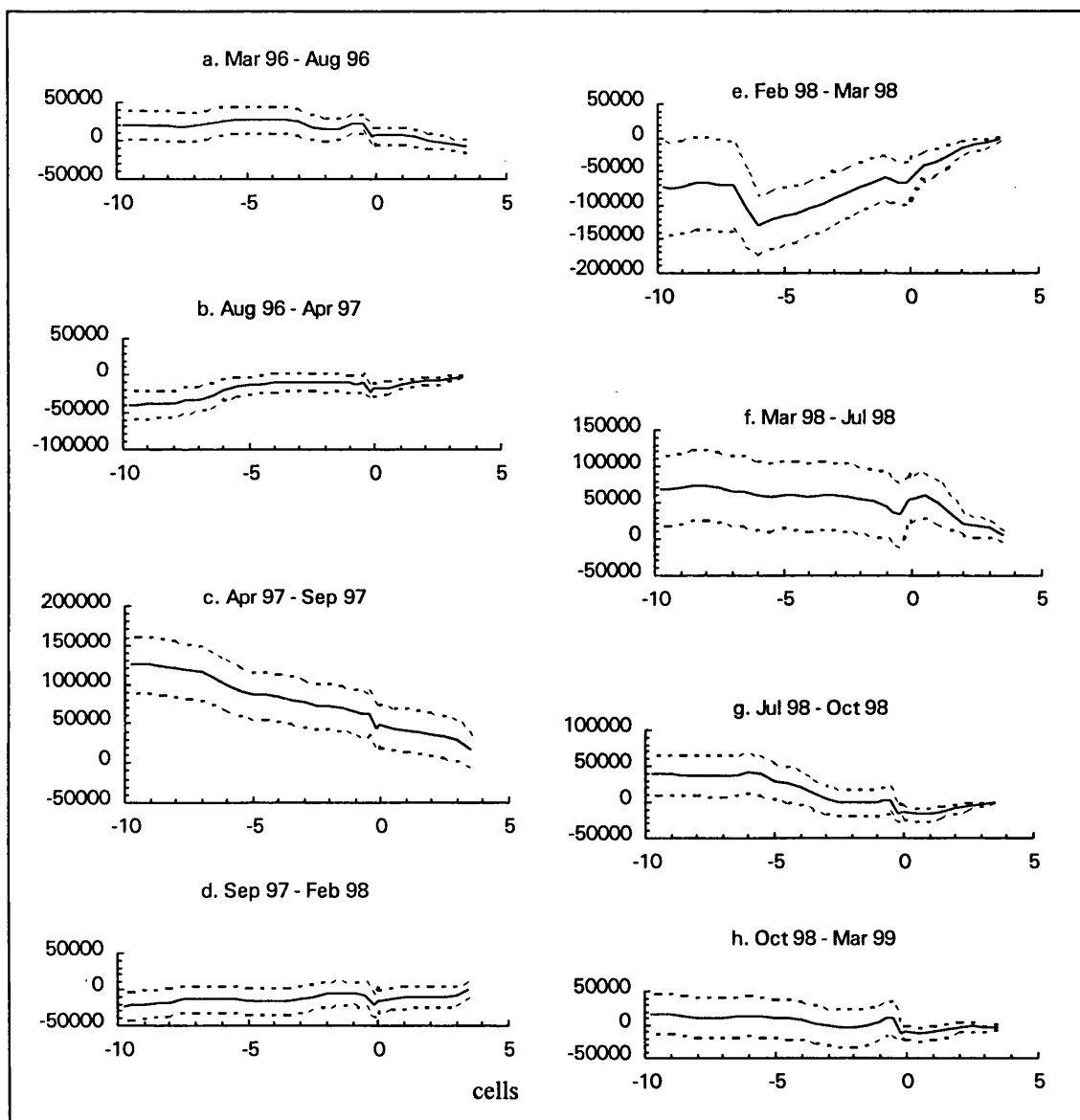


Figure 7.5: Estimated longshore gravel transport rates (in $\text{m}^3 \text{a}^{-1}$) for Spey Bay with error ranges for the periods a. March 1996 - August 1996 b. August 1996 - April 1997 c. April 1997 - September 1997 d. September 1997 - February 1998 e. February 1998 - March 1998 f. March 1998 - July 98 g. July 98 - October 1998 h. October 1998 - March 1999.

Cumulative errors in Q_0 are presented graphically in Figure 7.5 for each period. Errors increase in a down-drift direction and can be greater than the magnitude of the gravel transport rate (e.g. between October 1998 and March 1999, the estimated gravel transport rate past profile -9.8 is ca. $16\,000 \pm 30\,000 \text{ m}^3 \text{a}^{-1}$). Most of the uncertainty in the transport rates are due to errors in the beach storage change estimates (Section 6.6.2.4) which may

account for up to 92% of the error. Uncertainties due to the porosity estimate and grain size multiplier account for the remaining 5% and 3%, respectively.

Estimated gravel transport rates past profile -9.8 over the three years of survey give a mean rate of $17\,000\text{ m}^3\text{a}^{-1}$ (ranging from $126\,000 \pm 36\,000\text{ m}^3\text{a}^{-1}$ to $-72\,000 \pm 69\,000\text{ m}^3\text{a}^{-1}$) (Figures 7.4 and 7.5). However, we know that the volume of gravel that passes this downdrift boundary is negligible, as this marks the abrupt transition from a gravel to sand beach (Section 6.5). This implies that one of the assumptions made during the application of equation 2.10 is invalid and indicates something of the nature of coastal sediment transfers at Spey Bay.

The recorded accretion on the east flank of Spey Bay in several periods (Figure 6.21) questions the validity of the assumption of uni-directional westerly drift. For example, in the period July 1998 to October 1998 the beach east of profile +1 experienced a net gain of ca. $4\,600\text{ m}^3$ of gravel (Figure 6.21g). As no gravel enters the system from the east (Section 7.1.2.2), the only possible sources of this gravel are either longshore drift, derived from beach erosion, river input or the delta to the west, or an unknown amount of onshore movement of sediment which may have been stored in the nearshore zone during the previous period (Figure 6.21f).

The results suggest that although Spey Bay is subject to net westerly drift, transfers of sediment also occur in an easterly direction for short periods. Although wave refraction analysis predict a net westerly longshore drift of ca. $3\,000\text{ m}^3\text{a}^{-1}$ (Dobbie & Partners 1990) drift trends on a year-by-year basis differ in magnitude and direction demonstrating the sensitivity of the longshore drift system to the wave climate. Temporary phases of eastward drift have been documented (Section 3.5.1.4). With this in mind, the longshore rates of gravel transport presented in Figure 7.3 should be viewed with caution.

It is also concluded that on/off sediment exchanges may be occurring in the nearshore to account for the apparent loss of sediment in both the medium-term and short-term budget calculations. In spite of these caveats and uncertainties, the close consistency between the two budgets is encouraging and suggests that both methods are relatively robust methods of calculation of sediment exchange volumes on the upper beach. However, sediment transfers between the lower beach and nearshore and at the sub-aqueous delta are also important and may be of a similar magnitude to those occurring on the upper beach. From inspection of Table 7.2, and since there appears to be no loss of gravel alongshore, these may account for up to 70% of the total gravel transfers occurring at Spey Bay.

7.1.3 Comparison with other fluvial and coastal sediment transport rates

7.1.3.1 Fluvial transport rates

Transport rates in the lower River Spey vary from $41\,000 \pm 6\,000 \text{ m}^3 \text{ a}^{-1}$ (1993-1994) to $6\,000 \pm 4\,000 \text{ m}^3 \text{ a}^{-1}$ (1997-1999) (Figure 7.6). These are comparable to published rates calculated for wandering gravel bed rivers of similar scale to the Spey (e.g. the Chilliwack and Vedder Rivers in British Columbia (Martin and Church 1995; Ham 1996)) and compare well to the long-term Spey average of $8\,000 \text{ m}^3 \text{ a}^{-1}$ derived from river modelling (Dobbie & Partners 1990). The 30-210m wide Chilliwack River has similar morphological characteristics to the Spey with a mean annual flood ($Q_{2.33}$) of $313 \text{ m}^3 \text{ s}^{-1}$ and a threshold flow for gravel transport of $250 \text{ m}^3 \text{ s}^{-1}$ (Ham 1996). Transport rates were quantified using planimetric methods similar to those used herein and vary between $55\,000 \pm 10\,000 \text{ m}^3 \text{ a}^{-1}$ and $5\,000 \pm 2\,500 \text{ m}^3 \text{ a}^{-1}$ between 1952 and 1991 (Ham 1996).

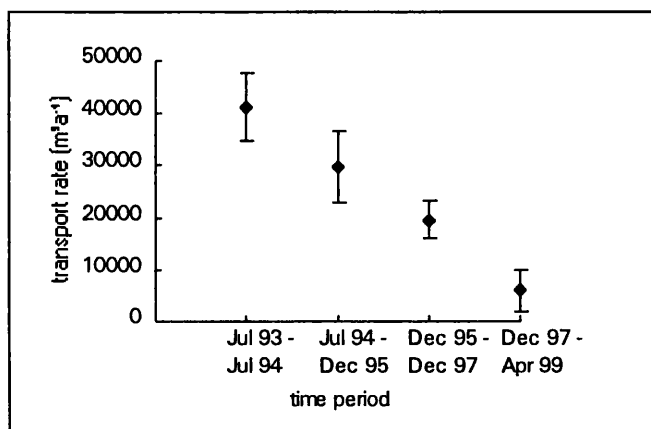


Figure 7.6: Variation in sediment transport rate for the lower River Spey (1993 - 1999) calculated using the morphological approach.

Martin and Church (1995) estimated a mean gravel transport rate of $36\,600 \pm 5\,600 \text{ m}^3 \text{ a}^{-1}$ between 1981-1990 for the Vedder River, the distal reach of the Chilliwack. In this case, reach storage changes were calculated from repeat cross-sections surveys at a mean downstream spacing of ca. 170m or approximately one channel width (Martin 1992). Error analysis of spacing suggested a cross-section spacing of between 250-300m (approximately twice the active width) was appropriate for estimating transport rates on the Vedder (Martin 1992) (Section 2.1.1.2).

The active channel width of the lower Spey varies between 50-350m (Figure 3.15) and the mean distance between depositional zones is 350m, therefore a cross-section spacing of 325m was considered appropriate (Section 4.1.1). Reach storage changes were estimated from repeat surveys of these sections (Section 5.2.1) and compared to estimates from

planimetric techniques (Figure 5.11). In some reaches the estimates are similar (reaches 4, 5 and 9), whereas in others they are vastly different both in magnitude and direction (reaches 3 and 10). This is because cross-section surveys often omit important zones of erosion or deposition (e.g. Figure 5.9) and thus fail to accurately estimate reach storage changes. These zones are included using planimetric methods making this a much more appropriate and reliable method to estimate transport rates on wide, wandering gravel bed rivers. A further limitation of cross-section surveys are that section locations are fixed and cannot evolve with the changing landform; an initially sensible location can rapidly become inappropriate due to channel change.

The unreliability of the Spey cross-sections to estimate reach storage change (Figure 5.11) suggests that a spacing of 325m (ca. twice the active channel width) is too wide to accurately calculate sediment transport rates on a river of this scale. Downstream spacing of ca. 100m *may* be more reliable on this river, although this would increase the field effort threefold with no guarantee of increased accuracy. Negative transport rates are estimated in most reaches (Appendix B), with the exception of the downstream reach where a rate of ca. $7\,800\text{ m}^3\text{a}^{-1}$ was estimated for the period 1997-1999. Interestingly, this rate is similar to the $6\,000 \pm 4\,000\text{ m}^3\text{a}^{-1}$ estimated for the same period from planimetric methods, although this may be coincidental.

Much wider spacing has been used to estimate transport rates on larger rivers (e.g. Griffiths 1979; McLean 1990, Section 2.1.1.2). Griffiths (1979) used sections spaced up to 1.6km to estimate transport rates of ca. $220\,000\text{ m}^3\text{a}^{-1}$ on the ca. 1km wide, braided Waimakariri River, New Zealand. In large channels, in which downstream patterns of erosion and deposition occur at larger scales, large cross-section spacing may be appropriate. Unfortunately, Griffiths (1979) gives no estimate of the uncertainty in his calculations. In the small and complex braided channels of proglacial streams, cross-section spacing must be small (e.g. Ferguson and Ashworth 1992; Goff and Ashmore 1994; Lane et al. 1994). Uncertainties must be quantified in all studies of transport rate estimates, as this work shows that inappropriate cross-section spacing can cause large errors in the estimation of reach storage change.

The planimetric method used in this study is considered more reliable but it is not without its limitations. Carson and Griffiths (1989) used the planimetric method to calculate volumes of erosion and deposition from consecutive aerial photography of the Waimakariri River. In all five measurement periods, volumes of erosion were substantially greater than volumes of deposition. Similar results were obtained on the Chilliwack River (Ham 1996)

and the Spey (Table 5.6), with erosion volumes exceeding deposition volumes for all periods. While this may represent the degradational nature of the channels, it may be a methodological problem. Erosion is generally localised and well-defined in braided and wandering rivers, whereas deposition covers a much larger area and is less easily mapped (Carson and Griffiths 1989) and will thus be under-estimated. In addition, planimetric methods do not account for any vertical changes on pre-existing bars. If volumes of deposition are under-represented transport rates will be over-estimated. Further refinement of the planimetric method is required to develop a technique for estimating this error and providing an appropriate correction factor.

The temporal density of re-survey (or re-mapping) affects morphological calculations mainly through changes in the probability of compensating erosion and deposition at any one location between surveys (Ashmore and Church 1999). As a result of this, transport estimates using the sediment budget approach are always a lower bound (Goff and Ashmore 1994; Lane et al. 1994). However the consistency between estimated gravel transport rates using planimetric methods and direct measurements of bedload transport on the lower Fraser River (McLean and Church 1999) indicate that on large rivers even with a time interval of 32 years this method does not introduce any significant negative bias.

The appropriate temporal density depends on the rate of channel change (Section 2.1.1.2). For large rivers, which experience change in response to floods, pre and post-flood surveys are required. The rates computed herein are calculated from planimetric changes computed from repeat geomorphological maps at 1-2 year intervals (Section 4.1.3). Mapping was carried out following major flood events and field evidence suggests that most morphological change takes place during the flood. Transport rates are given as the mean annual rates over the period between subsequent maps (Section 5.4.2), however it is likely that actual gravel transport rates are significantly higher during the flood event and may be negligible for most of the remaining period.

One of the major advantages of the sediment budget approach is that it quantifies the within reach variation in transport rates. For example, transport rates in the lower Spey increase rapidly downstream of reach 6 in all periods (Figure 5.19). This is where the river widens into a more braided reach with an increase in bare gravel bars and islands, resulting in increased morphological change (and hence increased transport rates) (see Figures 5.7 to 5.10).

7.1.3.2 Coastal longshore transport rate

The sediment continuity principle is often used to predict the areas of potential beach erosion, transport and deposition of sediment that result from spatial variations in the longshore energy flux (P_L , equation 2.13) (Greenwood and McGillivray 1978; Davidson-Arnott and Pollard 1980). Areas of potential erosion are associated with increasing values of P_L alongshore since more sediment would be transported out of the cell than into it. Conversely, deposition is indicated where P_L values decrease alongshore and where P_L values remain roughly constant a zone of transport without significant net erosion or deposition occurs (Davidson-Arnott and Pollard 1980). Many numerical models of shoreline change are based on this concept (Section 2.1.2) and use the variation in potential longshore transport rate to predict shoreline changes (e.g. Komar 1983, 1998).

However, few studies have reversed the sediment continuity principle using changes in beach cell volumes and equation 2.10 to estimate longshore transport rates. Drapeau and Mercier (1987) used this approach to estimate the sand sediment budget for a 70km long island, which was split into littoral cells ca. 500m wide and cell volume changes were assumed to be related to onshore-offshore processes. As balances ranged from $-24\,100\text{ m}^3\text{a}^{-1}$ to $+97\,600\text{ m}^3\text{a}^{-1}$ it was concluded that onshore-offshore exchanges were important, particularly in accreting cells (Drapeau and Mercier 1987). In another study, sand transport rates estimated using equation 2.10 for the Ebro delta coast, Spain (Figure 2.7), agreed well with those obtained using the CERC transport formulae and the average wave climate (Jimenez and Sanchez Arcilla 1993).

The approach used in Spey Bay was to quantify longshore rates of gravel transport using estimated beach cell volume changes and equation 2.10. All gravel (and most sand) sediment budgets calculate profile volume changes (in m^3/m) or cell volume changes (in m^3) and do not attempt any estimation of the longshore rate (e.g. Deruig and Louisse 1991; Hicks et al. 1999). This may be due to the lack of knowledge of appropriate boundary conditions, especially the offshore closure assumption, the directions of longshore transport and the need for an applicability of uni-directional longshore transport between cells.

The longshore gravel transport rates calculated for Spey Bay (Figures 7.3 and 7.4) perhaps highlight why this technique has seldom been used. Gravel transport past profile -9.8 should be zero (or close to zero) for all periods (Section 7.1.2.2) and not withstanding uncertainties in the estimated transport rates, the budget suggests that this is rarely the case

(Figure 7.4). It can thus be deduced that either the assumption of uni-directional drift is not entirely valid or onshore-offshore sediment exchanges are occurring. Nevertheless, the longshore transport gradient is indicative of shoreline erosion where it increases and beach accretion where it decreases. For most periods a slight decrease in the longshore transport rate around cells -1.5 to -2.5 occurs due to beach accretion in these cells (Figure 7.3).

In summary, this technique should be used with caution when estimating longshore gravel transport rates, unless all onshore-offshore sediment exchanges can be quantified and longshore drift directions are well established. Both tasks are extremely difficult to fully accomplish. However, if errors are minimised and the technique works well, this may be the best way of learning about onshore-offshore exchanges.

7.2 Long-term budget changes (10 000 years)

The short-term (1996-1999) and medium-term (1870-1970) budgets for Spey Bay are both negative, indicating a system that has been erosional for at least the last century (Section 7.1). Notwithstanding the uncertainties inherent in the volume change estimates (Section 7.1.2.3) the map, morphological and field data provide substantial evidence that the Spey Bay system over the recent period is predominantly erosional. This section discusses the nature of sediment supply throughout the Holocene (Section 3.2), its effect on the budget and morphology of Spey Bay and seeks to place the current erosion into a longer-term context.

The rise in relative sea-level at the peak of the Holocene transgression (ca. 6500 BP) brought sediment onshore from the inner shelf and, combined with high sediment discharges from the Spey, created a sediment-rich coastal environment in Spey Bay (Section 3.2). By estimating the volumes of sediment contributed to the Spey system via fluvial transport during the Holocene and matching the volume against sediment contained within the gravel ridge systems at Spey Bay, it is possible to estimate a palaeo-sediment budget for the system. Comber (1993) attempted to do this by calculating the volume of sediment removed from the lower Strathspey terrace sequence, based upon reconstruction of the terrace fragments identified by Peacock et al. (1968) (Figure 3.8). The total volume removed since the formation of the Lateglacial terrace surface was estimated to be $3.35 \times 10^8 \text{ m}^3$ (Comber 1993). Borehole data from the lower Spey terraces suggests that the mean proportion of gravel in the terrace sediment is 88%, therefore $2.94 \times 10^8 \text{ m}^3$ of gravel was removed from the terraces during the Holocene. Assuming constant rates throughout the Holocene and correcting for porosity this equates to sediment removal rates of ca. 200 000

m^3a^{-1} . This is a minimum value as Comber (1993) ignored the supply of sediment from terrace erosion further upstream in the Spey system.

The volume of gravel contained in the raised gravel strandplain at Spey Bay (Figure 3.5, Section 3.2) was estimated to be $5.4 \times 10^7 \text{ m}^3$ (Table 7.3) based on the surficial area of gravel multiplied by a depth of 6 m, this being the estimate used by Comber et al. (1994) of the mean depth of gravel at Culbin.

	Area (m^2)	Volume (m^3)
West Spey Bay	7.23×10^6	4.34×10^7
East Spey Bay	1.78×10^6	1.06×10^7
Total	9.01×10^6	5.4×10^7

Table 7.3 : Estimated volume of gravel in the raised gravel strandplain at Spey Bay

Assuming a constant rate of sediment deposition over the Holocene, the rate of sediment accumulation in the ridges of Spey Bay is of the order of $5\,400 \text{ m}^3\text{a}^{-1}$ over the Holocene. Using values for beach porosity of 0.3 and assuming the beach is composed of 95% gravel this represents gravel accumulation of the order of $3\,500 \text{ m}^3\text{a}^{-1}$. The short-term and medium-term budgets reveal large losses of gravel from the beaches of Spey Bay, amounting to around $5\,300 \text{ m}^3\text{a}^{-1}$ and $3\,900 \text{ m}^3\text{a}^{-1}$ over the periods 1996-1999 and 1870-1970, respectively (Tables 7.1 and 7.2). If the estimates are accurate, the budget has switched from one of net accretion at some time during the Holocene to one of net erosion and sediment deficit in the late Holocene. This has important implications for beach development.

Interactions between relative sea level (RSL) and sediment availability are central to the functioning of the coastal system (e.g. Carter et al 1987; Roy et al. 1994) and it has been argued that a mid-Holocene switch from sediment surplus to deficit at about 6.5 ka BP (Carter 1988) led to erosional shoreline tendencies irrespective of RSL sense in Europe and elsewhere (Hansom 2000). An important repercussion of the mid-Holocene decline in sediment supply was the re-organisation of coastal sediment into progressively smaller coastal cell and sub-cells that are dominated by the internal re-organisation processes of erosion and deposition (Hansom 2000). The morphology and sediment budget results from Spey Bay fit this pattern.

Under declining sediment supply, the present beach ridge at Spey Bay is unlikely to represent the next sequential addition to the seaward edge of progradational ridges (as occurred in the mid-Holocene under positive sediment budgets) and is more likely to be a

product of the reworking of older ridges to the landward. As the system is effectively starved of sediment, it begins reworking previously deposited beach sediment, so beach accretion is predominantly fuelled by updrift beach erosion or episodic fluvial input. The future evolution of Spey Bay is likely to be one of continual internal re-organisation within the erosional cell, with volumes of sediment added as point sources at the Speymouth delta and erosional sites. The decline of sediment supplied from offshore and the switch to a negative sediment budget in the late-Holocene (Hansom 2000) is likely to lead to an increase in the relative importance of gravels sourced from the Spey for beach budgets. However, with gravel inputs declining through the Holocene (Young 1977; Maizels 1988) and short-term temporal variability causing rates to vary from $41\,000$ to $6\,000\text{ m}^3\text{a}^{-1}$ between 1993-1999 (Figure 7.6) this has major implications for coastal stability.

7.3 Processes and implications for system operation

The net changes derived from three years of morphological survey indicate that the contemporary Spey Bay system operates in a similar way as it has done for at least the last century (compare Figures 7.1 and 7.2). Most of Spey Bay is erosional, with the exception of the stretch of beach around Boar's Head Rock where long-term gravel accretion has taken place as the gravel front migrates. The sand beach at Lossiemouth is erosional over the medium-term (Figure 7.1) and this is corroborated with results from short-term morphological data (Figure 7.2).

However, the morphological surveys also indicate that the short-term phases of coastal erosion and accretion are neither spatially nor temporally uniform. For example, the eastern part of Spey Bay undergoes phases of erosion and accretion at different times (Figure 6.21) within a general erosional trend. In addition, the location of zones of accretion and erosion vary spatially through time. While this may be attributed to seasonal factors and the on/offshore movement of sediment, there are likely to be other factors driving the sediment exchanges identified at Spey Bay. These are discussed below.

7.3.1 Fluvial sediment supply

The rate of sediment input from the River Spey to the coast during the study period was variable and decreased between 1993 and 1999 (Figure 7.6). This, together with the operation of the delta (Section 5.5), has implications for coastal stability and may account for some of the temporal and spatial variability observed on the coast. For example, fluvial transport rates fell to $6\,000 \pm 4\,000\text{ m}^3\text{a}^{-1}$ between December 1997 and April 1999 and given that the Speymouth delta increased in volume by ca. $4\,000 \pm 200\text{ m}^3\text{a}^{-1}$ (Section

5.5.2) only ca. 2 000 m³a⁻¹ of sediment was available to nourish the coast. The high rates of erosion of the Spey Bay beaches recorded over this period (Section 6.6.2) may therefore be a result of reduced sediment supply.

Reasons for the variability and overall reduction in fluvial transport rates were discussed in Section 5.4.3. Although there is a strong relationship between the number of flow events over a given threshold (in this case $Q_{0.5}$) and the sediment transport rate for any given period (Figure 5.21) transport rates are not directly related to flow conditions. Other factors such as sediment supply, sediment accessibility and channel stability were shown to be important controls of bedload transport rates in wandering gravel-bed rivers. Access to sediment is a critical control on transport rates. The stability of the channel and bars prior to a flood influences subsequent transport rates. For example, the channel avulsion in January 1994 created large areas of bare gravel which became available for transport in subsequent events, explaining the high transport rates in the periods immediately succeeding the avulsion (Figure 7.6).

Reservoir theory (Section 2.1.1.1) can be used to describe the sediment exchange processes observed in the lower Spey. Fluvial sediment can be classified into four reservoirs: active, semi-active, inactive and stable (Figure 2.1) based on the relative mobility of the sediment. Sediment can only exit (or be absorbed) from the channel via the active reservoir (Figure 2.2) and can ‘move’ between reservoirs via dynamic exchanges (where the sediment itself moves) and static exchanges (where reservoir boundaries move) (Hoey 1996). Static exchanges of sediment between reservoirs play an important role in the lower Spey. For example, “inactive” or “stable” sediment in the wooded floodplain prior to the channel avulsion in January 1994 moved rapidly to the “semi-active” state after the avulsion as the new avulsed channel was now much closer to the stored sediment. The sediment itself did not move but its transport potential was greatly increased. The “active” unvegetated gravel bars (e.g. bar b) in the reach downstream of the viaduct in 1995 progressively moved to the “semi-active” state as vegetation established. The decrease in the amount of “active” sediment, as vegetation increased on the bars and islands between 1995 and 1999, may have contributed to the observed decrease in transport rates during the study period (Figure 7.6). Access to available sediment is therefore a critical factor influencing bedload transport rates. Flood events play an important role in increasing the amount of potentially active sediment, either by channel avulsion or vegetation stripping during over-bank flows.

There is evidence that channel change is controlled by variations in upstream sediment supply (Lane et al. 1996). Excess sediment in a given reach promotes morphological

change and channel instability. The 3-year flood which occurred in July 1997 caused less extensive morphological change in the lower reaches of the Spey than previous floods of similar magnitude, and thus had less effect on transport rates (Section 5.4.3). This may be related to the increased stability of bar deposits, as vegetation established, but it is likely that natural removal of sediment from upstream to infill the former Essil channel was a contributory factor. Over 25 000 m³ of sediment was deposited in the former Essil channel during this flood (Section 5.4.2). This sediment essentially became trapped (stored) in the upper reaches and was unable to create morphological change and instability further down the system. This may account for lower transport rates than expected in this period, given the magnitude of the flood peak (Figure 5.22).

Results from the Fraser River in Canada corroborate these findings. McLean and Church (1999) found that transport rates varied dramatically for a given flow condition according to the availability of sediment for transport. Even though potentially available sediment is present in the beds and banks of the channel, its delivery to the flow depends upon morphological and structural characteristics of the sediment pile. As with the Spey, sediment is mobilised according to the recent local history of erosion, sediment deposition and channel adjustment (McLean and Church 1999). Current theoretical approaches to sediment transport fail to account for such complexities and further research is required.

7.3.2 Processes at the fluvial-coastal interface

The process of westerly spit growth at the Speymouth delta has implications for the storage and transfer of sediment between the fluvial and coastal system. Work at the Rakaia river mouth (Kirk 1991) and other rivers in New Zealand (Single 1999, pers. comm.) describe a model of spit growth and breaching which is of relevance to the Spey system. During low river flows, long narrow spits extend across many New Zealand river mouths, as a result of the prevailing northerly longshore drift, diverting the exits accordingly. During higher discharges, the spits are breached by the river and a “slug” of the drift sediment temporarily stored in the spits is bypassed to the downdrift coast (Kirk 1991). Pulses of sediment moving north along the Canterbury coast, New Zealand are thought to be related to the release and storage of slugs of sediment at river mouths (Section 2.2.2).

A similar process is observed at the Speymouth delta (Figure 5.27, Section 5.5.4) which has important implications for the downdrift coast. The study period was characterised by a period of westerly extension of a gravel spit complex extending at ca. 150 m a⁻¹ across the river mouth. A migrating zone of accretion was observed in the lee of the spit, where a

series of low recurving gravel ridges became attached to the lower foreshore (Plate 5.2) and migrated alongshore at similar rates as spit extension. Beach accretion recorded in cell -0.5 to -0.8 up to July 1998 (Figure 6.21) documents the gain of sediment as successive ridges weld onto the lower foreshore. The loss of sediment in this cell during the last two beach surveys (Figures 6.21g and h) provides evidence for updrift erosion of the accretionary lobe that fuels its downdrift migration (Figure 5.27). The erosional bight which precedes the migrating accretionary lobe is corroborated by beach profile data which record a loss of sediment in cell -0.8 to -1 over most of the study (Figure 6.21). Accretion in this cell recorded during the last survey (Figure 6.21h) reflects the westward migration of the accretionary lobe. Interactions at the river mouth result in the slow movement of pulses of erosion and deposition along the coast and can cause significant variations in the longshore transport rate that are most notable close to the mouth but which are propagated downdrift.

Similar processes have been documented on both gravel (Kirk 1991) and sand beaches (Hicks and Inman 1987; Hicks et al. 1999) where the incorporation of fluvial sediment results in erosion/accretion cycles along the downdrift coast (Section 2.2.2). These have been identified at distances up to 3-4km alongshore from the river mouth (Hicks et al. 1999).

7.3.3 Temporal and spatial variation in longshore gravel transport

Recent work has identified bed waves (or “slugs” of sediment) in the fluvial system which cause temporal and spatial variation in bedload during quasi-stable flow (Nicholas et al. 1995). Pulses of sediment moving along the coast have also been documented in response to variations in river sediment supply (Gibb and Adams 1982), river mouth processes (Hicks and Inman 1987; Neale 1987; Kirk 1991), beach feeding (Grove et al. 1987), storm events (e.g. Matthews 1980; Todd 1989; Bray et al. 1996; Brunsden 1999; Single and Hemmingsen 2000) and inlet processes (Hicks et al. 1999). At Spey Bay there is evidence of similar pulses (or “slugs”) that move along the coast.

The storm event of March 1998 resulted in alternating zones of erosion and deposition along the coast (Section 6.6.6.2) and profile response varied depending on alongshore location. Some profiles experienced upper beach erosion and sediment loss (e.g. profile -0.8), while others experienced upper beach accretion and net sediment gain (e.g. profile -3) (Table 7.4 and Figure 7.7). The distance between erosional and accretionary sections following this storm event is approximately 2-3km. Morphological evidence of storm overwash and lobes of gravel conform to this pattern of phased erosion and accretion in response to north-easterly waves and a westerly longshore drift during the storm.

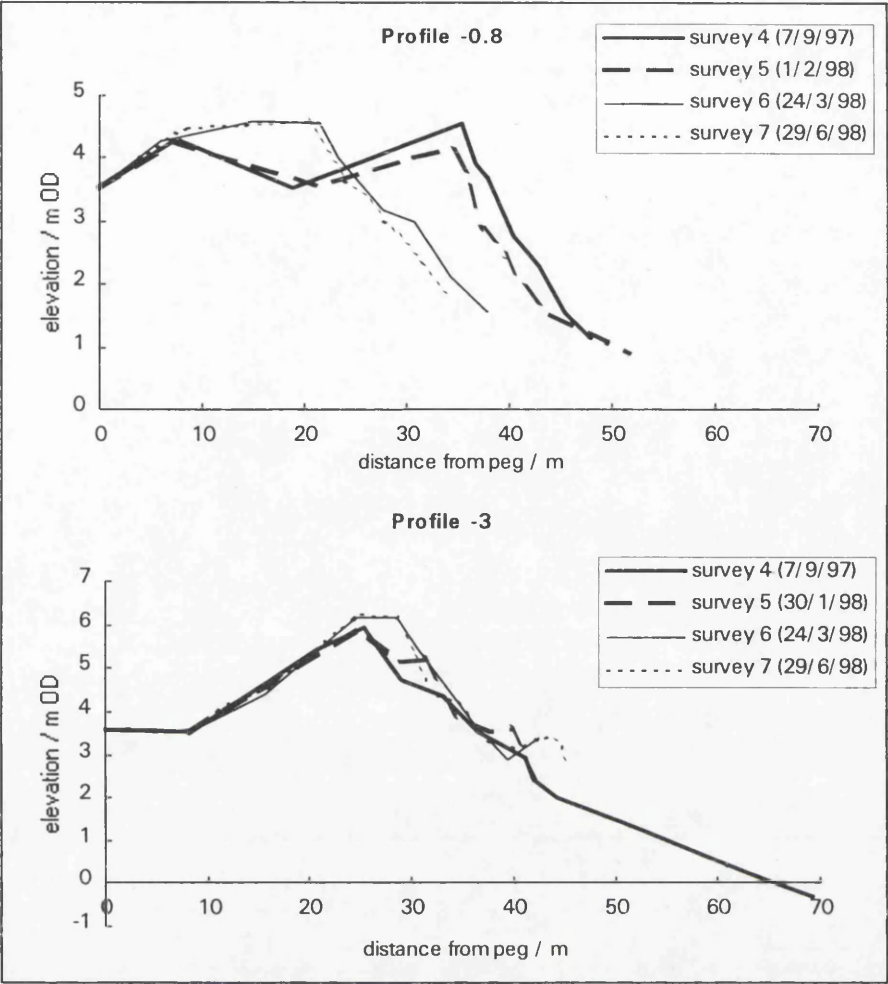


Figure 7.7: Morphological response of the coarse-clastic beach at Spey Bay to the storm event of 1/3/98 (survey 5 to survey 6). The updrift beach (profile -0.8) experienced major erosion, with crestal overwash and retreat of ca. 13m, while the response at the downdrift beach was one of upper beach accretion and crestal advance. The profiles are ca. 2km apart.

Profiles	Morphological Response	Movement of Crest
+4 to +1.5	erosion	landward
+1 to -0.1	accretion	seaward
-0.5 to -1.5	major erosion	landward
-2 to -3.5	accretion	seaward
-4	negligible	none
-4.5 to -6	erosion	minor landward retreat
-6.5	accretion	seaward
-7 to -8.5	erosion	landward
-9	minor accretion	seaward
-9.5 - 9.75	erosion	retreat, but substantial increase in crest altitude

Table 7.4: Summary of the coarse-clastic beach profile response to the storm of 1st March 1998 (see Appendix F). Note the alternating zones of erosion and accretion along the coast.

It is suggested that slugs of sediment sourced from updrift erosion move downdrift in response to storm events by a given distance to fuel accretion on downdrift beaches. The magnitude of each storm influences the size and transfer distance of the sediment slug. The available evidence suggests the March 1998 storm has a transfer distance (i.e. the distance between erosional and accretional beach sections) of approximately 3km. Patterns of alternating sections of volumetric gain and loss along the beach during other periods (e.g. March 1996 to August 1996, September 1997 to February 1998 and October 1998 to March 1999, Figures 6.21a, d and h) provide further evidence of the alongshore movement of slugs. This concept is supported by work on gravel beaches elsewhere (Matthews 1980; Bray et al. 1996) and is emerging as an important, and underestimated, process in recent work (Brunsden 1999; Single and Hemmingsen 2000).

Tracing experiments at Shoreham beach, England, have found that gravel moves alongshore in rapid bursts in response to high energy events, with longshore movement of up to 600m recorded during one storm (Bray et al. 1996). Matthews (1980) work on New Zealand beaches also reveals that gravel does not move alongshore coherently, but moves as small slugs which are separated from the main gravel body and moved rapidly alongshore. The small slugs (of the order of $1\text{--}10\text{ m}^3$) move between 10-50 m per day (Matthews 1980). The longshore pulse recorded at Spey Bay in response to the March 1998 storm event involved larger volumes of sediment (e.g. $1\,500\text{ m}^3$ of sediment was released from the 700m stretch of beach between profile -0.8 and -1.5) and longer transport distances (ca. 2-3km) than previous findings, but provides further evidence for the pulsed nature of longshore gravel transport on coarse-clastic beaches. Smaller slugs may be present at Spey Bay that the sampling density was unable to detect.

Evidence exists of slugs of gravel moving west and east along Chesil Beach, Dorset (Brunsden 1999). It is argued that small slugs move over bigger ones and along different storm ridges, forming and reforming as conditions dictate. This process leaves remnants of slugs at different beach levels according to the severity and sequence of storms. While this model is attractive, it is difficult to demonstrate with conclusive field evidence (Brunsden 1999). However, there is some field evidence from the mixed sand and gravel beaches of New Zealand which support this model (e.g. Neale 1987; Todd 1989; Single and Hemmingsen 2000). Short-term data from the South Canterbury coast suggests that cells of erosion or accretion migrate downdrift, with different parts of the beach experiencing erosion or accretion at different times within a long-term trend of erosion (Todd 1989). Neale (1987) and Single and Hemmingson (2000) identified a periodic northward

movement of slugs of sediment along the South Canterbury coast, reflecting the irregularity of the sediment supply and the episodic nature of southerly storm events. These slugs can increase the volume of the beach by over $100 \text{ m}^3 \text{ m}^{-1}$ and travel downdrift at an average rate of 1.3 km/yr (Single and Hemmingson 2000).

Morphological surveys of Spey Bay show large temporal and spatial variability in altitude at different levels of the beach (Appendix F) which provide evidence of slugs of different sizes moving varying transport distances. For example, following the March 1998 storm event a large amount of gravel became attached to the upper beach at profile -3.0, increasing the height of the storm ridge by 0.3 m (Figure 7.7) and increasing the beach volume by ca. $5 \text{ m}^3 \text{ m}^{-1}$ (Figure 6.18). This slug of sediment remained in storage on the upper beach at profile -3.0 throughout the following year, as no storms of sufficient magnitude occurred to remove it. However, variations in the beach-face at profile -3.0, with tidal and storm ridges forming and reforming at lower altitudes than the storm crest, provide evidence of smaller slugs of gravel (requiring less intense storms) moving alongshore. These move mainly west along the coast of Spey Bay, depending on wave approach angles. Major storm events at Spey Bay occur during north-easterly swell, due to the increased fetch in this direction (Section 3.5.1.5), so the largest slugs (in terms of both magnitude and transport distance) will move west along the beach. Morphological data supports this.

Longshore pulses in sediment transport also reflect variations in longshore sediment supply. For example, beach recharge was carried out at Spey Bay just prior to the first beach survey (Section 6.3). Approximately $15\,000 \text{ m}^3$ of sediment was removed from the lower beach in the vicinity of profile -0.5 and deposited on the upper foreshore around profile -1.0 in late March 1996. In August 1996, beach accretion was observed to the west of the recharge area, with increases in upper beach volumes of around $5\,000 \text{ m}^3$ recorded along the 1 km (ca. $5 \text{ m}^3 \text{ m}^{-1}$) stretch of beach between profile -1.5 and -2.5 (Figure 6.21a). The morphology was characterised by a wide upper beach with many ridges and accretionary cusps, indicating upper beach storage of sediment and remnants of depositional slugs (e.g. Plate 6.8). As the initial recharge area recorded a loss of sediment during this time (Figure 6.21a) it is postulated that the recharge sediment has merely moved downdrift fuelling accretion on the downdrift beaches. Whether this occurred during one storm event or gradually, as a series of slugs over the 5 months, is unknown. The downdrift wavelength of this sediment pulse is ca. 0.5 km .

7.3.4 Westerly migration of the coarse-clastic beach front

At the west end of Spey Bay, the coarse-clastic beach front has been migrating westwards at ca. 33 m a^{-1} since 1870, progressively replacing the sand beach at Lossiemouth, although there is evidence that the rate has slowed recently to ca. 22 m a^{-1} between 1994-1998 (Table 6.1). Similar migration of gravel fronts, at rates of up to 420 m a^{-1} , have been recorded on New Zealand beaches (Matthews 1980). Evidence collated during surveys at the westerly extent of the coarse-clastic beach (profile -9.8) suggests that the process of westerly migration is temporally variable depending on wave and sediment supply conditions and may be related to the occurrence and arrival of sediment slugs. The visible extent of gravel varied between 0-100m west of profile -9.8 between March 1996 and March 1999 (Section 6.3). After periods of westerly winds, the gravel front was often obscured by a veneer of wind blown sand from the east. Extension of the coarse-clastic beach occurred during periods of westerly longshore transport, when generally small spherical gravels accumulate at the back of the sand beach beneath the eroding dune. Gravel accumulation progressively forms a low recurving ridge at the back of the beach and during subsequent periods of westerly drift, a second gravel ridge forms to the seaward, thus increasing the width of the coarse-clastic beach. This process continues and the beach builds seawards and westwards as more gravel ridges become welded onto the upper beach. Morphological changes at profile -9.75 document this process, with the gravel to sand boundary on the lower foreshore moving progressively seawards at ca. 10 m a^{-1} during the study (Section 6.3). Storms rework the low recurving ridges creating higher, more prominent, beach ridges.

The coarse-clastic beach west of profile -5.5 has up to five low recurving gravel ridges landward of the main active beach crest. These ridges represent the process of westerly migration of the gravel front; the most landward ridge is the oldest and at the time of initial deposition would represent the westerly limit of the coarse-clastic beach. This is the lowest in altitude and consists of generally smaller, more spherical gravel, than the recurving ridges further seawards. Some of the more seaward ridges are truncated by the present beach (e.g. profile -5.5) suggesting long term erosion and planimetric readjustment of this part of Spey Bay. Updrift erosion of the present beach is fuelling downdrift accretion to form new recurving ridges at the migrating westerly gravel front. Between 1996 and 1999 some $4\,250 \text{ m}^3$ of sediment accumulated at the westerly extent of the coarse clastic beach (profiles -9 to -9.8) (Figure 6.22).

Immediately west of the migrating gravel front, the beach is characterised by a wide intertidal sand beach backed by ca. 6m high erosional sand dunes (Plate 3.6). Severe erosion

occurs in the cell downdrift of the gravel front (-9.8 to -10.5), which recorded a net loss of ca. $-8\,000\text{ m}^3$ of sediment between 1996 and 1999 (Figure 6.22). Gravel emplacement at the back of the sand beach serves to protect the dune from wave undercutting and further erosion. Thus, dune erosion ceases as the gravel front migrates westwards and the dune-face stabilises and is colonised by vegetation. This is exemplified by the high erosion rates recorded along the stretch of beach west of profile -8.5 between 1870 and 1970 (Table 7.1) which was unprotected at this time (Figure 6.1) and the accretion that has occurred since (Table 7.2). Progressive replacement of the sand beach by the coarse-clastic beach creates a different shoreline orientation and the upper beach becomes higher and steeper (Plate 6.1).

7.4 Conceptual model of gravel sediment transfers from rivers to beaches

A conceptual model of gravel sediment transfers from rivers to beaches (Figure 7.8) is developed based on the Spey Bay data. Sediment transfers of gravel sized sediment from rivers to the coast are temporally variable and depend not only on the magnitude and frequency of flooding, but also on sediment accessibility prior to the flood (which is related to flood sequencing) and upstream sediment supply (Section 7.3.1). These fluctuations in fluvial sediment transfers result in pulses of sediment being supplied to the mouth which affect the dynamic balance of the coast (cf. Gibb and Adams 1982).

Processes at the river mouth exert a major influence on the release of fluvial sediment to the downdrift coast. Spit growth and westerly deflection of the river mouth can result in the downdrift migration of zones of beach accretion and erosion. Hicks et al. (1999) show that processes at an ebb tide delta can induce erosion/accretion cycles on sand beaches up to 3-4km alongshore from the inlet. Interactions at Speymouth have a direct effect on the coast at least up to 1.5km alongshore but indirect effects may be felt well beyond. In certain cases (e.g. May 1997-May 1998) sediment accretion at the delta can account for almost all the fluvial input to the coast. This complete but temporary storage at the delta serves to induce erosion on downdrift beaches, although later storm events may release the stored delta sediment to the downdrift coast as an accretionary pulse. Similar processes have been observed at a range of New Zealand river mouths including those of similar scale and discharges as the Spey (Kirk 1991).

Longshore changes in beach cell volumes are temporally and spatially variable, with most cells undergoing phases of erosion/accretion as part of a long term erosional trend, depending on location alongshore as internal re-organisation within the sediment cell proceeds. Morphological data indicates that gravel transport occurs in pulses (of different

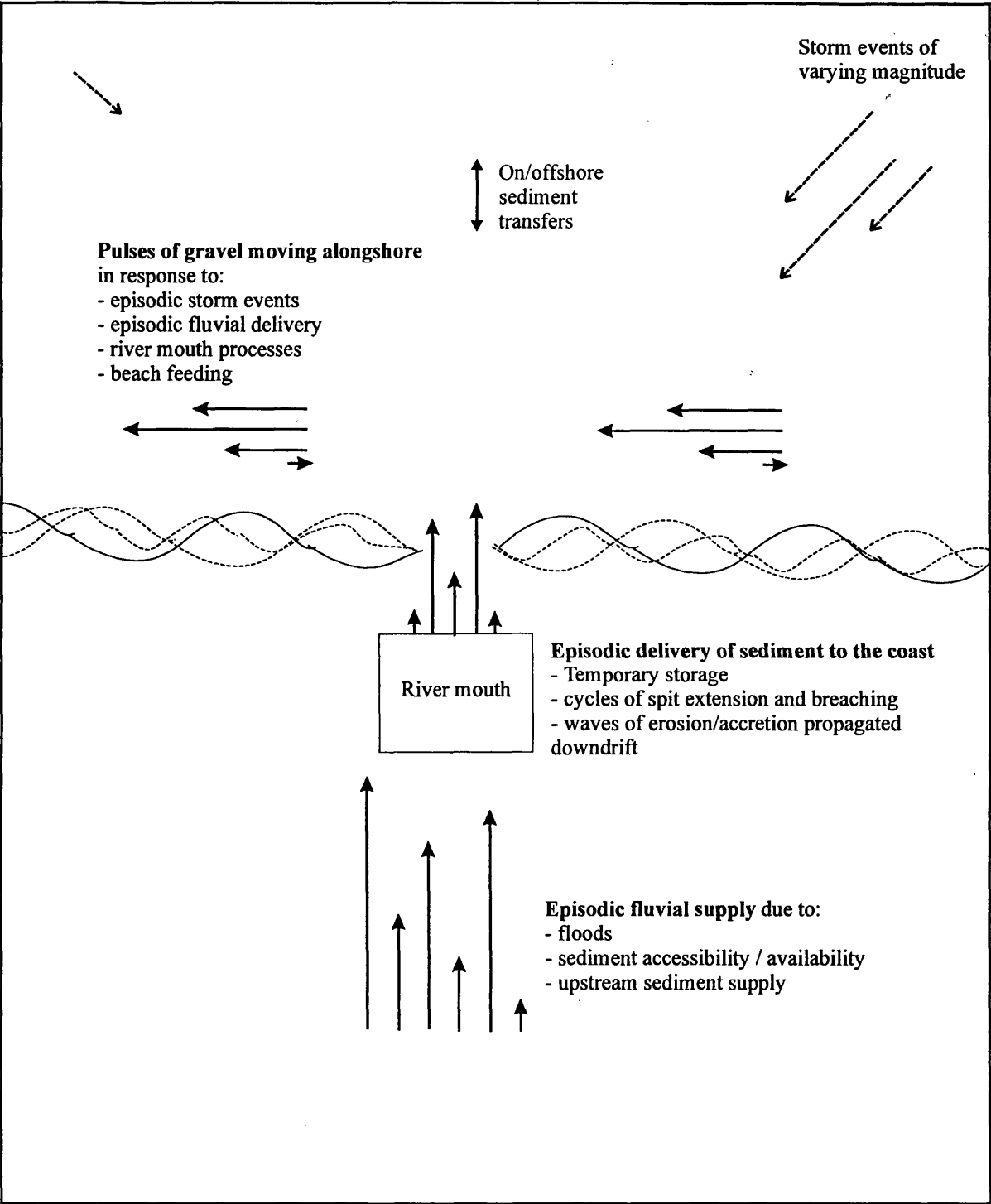


Figure 7.8: Conceptual model of gravel sediment transfers from river to beaches

magnitudes and in mainly westerly directions) related to storm events, river mouth processes, fluvial sediment supply and beach recharge. This has important implications for future beach management as zones of erosion (or “problem” zones) are not static and vary through time, probably systematically. An erosional cell may subsequently record accretion as a sediment pulse moves alongshore. With larger, and possibly more detailed, data-sets (both in time and space) it may be possible to identify volumes and periods of slugs associated with storms of particular magnitude. The conceptual model could thus be developed into a predictive model, to allow identification of zones of erosion and accretion for specific storm events.

The Spey coastline undergoes net westerly drift and although drift occurs in both directions depending on the wave climate, easterly drift is very small. The largest pulses of sediment moving along the coast are likely to be in a westerly direction, as the severest storms are from the north-east (Section 3.5.1.2). The zones of beach accretion which are created during major storm events (e.g. March 1998) have longer residence times on the upper beach than smaller pulses, as they require a storm of similar magnitude to initiate movement. Thus the response of a given stretch of beach is not only dependent on the magnitude of the storm event and tidal height but also on the antecedent beach condition (cf. Single and Hemmingsen 2000). This dependence on antecedent beach conditions suggest that the Spey Bay profiles display something of the Markovian characteristics found on mixed-sediment beaches elsewhere (Sonu and James 1973; Mason and Hansom 1989). Markovian models suggest that the present beach state depends on some probability function of the immediately preceding state, but not on any state before that, implying that no long-term memory exists in the beach system (Mason and Hansom 1989).

The conceptual model (Figure 7.8) is applicable to gravel fluvial-coastal systems in general. This study has highlighted the spatially and temporally variable nature of sediment transfers in fluvial and coastal systems and highlights some of the complex issues that numerical simulations or theoretical equations need to account for when used to quantify such transfers. These often assume constant operation of processes and do not take account of the inherent variability in the system which may be related to the variable morphological and structural characteristics of the sediment. The time-scale of study is important, as over longer periods such fluctuations may average out with the trends still being preserved.

7.5 Methodological issues

The conceptual model presented above attempts to capture the general sense of gravel sediment transport from rivers to beaches and is transferable to other fluvial-coastal gravel cells. It highlights the inherent linkages between fluvial, deltaic and coastal systems and demonstrates the spatial and temporal variability of sediment transfers in response to river floods, sediment availability, river mouth processes and coastal storms. The conceptual model may form the basis of a predictive model, whereby the volumes and periodicities of pulses of fluvial sediment input and alongshore migrating slugs of sediment can be related to the magnitude and frequencies of both river floods and coastal storms. This would require a much denser survey network of beach and river profiles (both in time and space) in order to identify smaller pulses/slugs with shorter travel distances and will be problematic logistically in large systems. Future work could perhaps concentrate on the ca. 3-4km stretch of coast downdrift of Speymouth or develop a similar approach on a smaller system or physical model. Nevertheless, the present study has clearly identified spatial and temporal variations in river to beach sediment transfers and sets the agenda for future research and modelling. Appropriate spacing of beach and river profiles and an adequate definition of the ‘closure depth’ may be critical to the success of future research.

7.5.1 Profile spacing

The downstream spacing of cross-sections required to accurately quantify storage changes in a given reach (δS) varies depending on the scale of the river (Sections 2.1.1.2 and 7.1.3.1). Cross-section spacing is often related to the channel width (cf. Martin and Church 1995) or the transfer distance (i.e. step length) between deposition (or erosion) zones (Carson and Griffiths 1987). A mean downstream spacing of ca. 2 times the active channel width (and approximately equal to the transfer distance) was used in this study (Section 5.2.1). However, this proved unreliable to estimate reach storage changes (Section 5.2.3), indicating that a much denser cross-section spacing is required if this method is to be successfully applied to wandering gravel-bed rivers of similar scale to the Spey. Further research is required to determine a system for determining cross-section spacing according to river scale. Furthermore, all studies that use cross-section surveys to estimate sediment transport rates must include a formal error analysis of the derived transport rates. This study has shown that errors can be substantial (Appendix E) and corroborates Martin and Church’s (1995) results. The planimetric approach (cf. Ham 1996) is considered a more reliable method for rivers of this scale.

Previous studies which have estimated beach cell volume changes from successive beach profiles have omitted analysis of the errors arising from beach profile spacing (e.g. Comber 1993; Foster et al 1994; Hicks et al 1999). This study has determined that the longshore spacing of beach profiles can significantly influence the accuracy of beach cell storage change estimates (Section 4.6.2.8). Errors can be up to $\pm 118\%$ of the estimated volume change at a 500m spacing (Figure 4.18), although the percentage error declines with the magnitude of the actual change. Large volumetric changes are likely to be associated with smaller percentage errors. Formal error analysis of the uncertainty in cell volume changes estimated from beach profiles should be included in all future sediment budget studies.

7.5.2 Closure depth

Sediment budgeting has been described as one of the most useful concepts in coastal research and management, the development of which should be the goal of all coastal-management programs (Komar 1996). International examples abound in the literature of the analysis of beach sediment budgets and volumes (Deruig and Louisse 1991; Lacey and Peck 1998; Hicks et al. 1999) and their applicability for coastal management (e.g. Bray et al. 1995).

However, most sediment budgets are constructed using data restricted to mean sea-level or just below low water mark (e.g. Lacey and Peck 1998; Hicks et al. 1999), when ideally all profiles should be surveyed to the closure depth on all occasions (cf. Foster et al. 1994). Many studies either extend surveys out to the depth of closure at lower frequencies than the subaerial surveys, although this limits the reliability of the volume changes in the lower shoreface (Deruig and Louisse 1991), or extrapolate surveyed beach profiles out to a common depth of closure (Pierce 1997) or MSL (Lacey and Peck 1998). The lack of real data on the lower shoreface can be a problem as the magnitude of volume changes here may be 2-3 times those on the upper beach and swash zone (Deruig and Louisse 1991; Foster et al. 1994).

This study highlights the problems of extrapolating surveyed profiles out to a common depth for volume change analysis (Section 6.6.1). Volume changes based on extrapolated profiles are generally greater than those based on the actual surveyed profiles and sometimes record changes in the opposite direction (Figure 6.18). The differences between the original and extrapolated data-sets are heightened when profile changes are averaged over the 500m beach cell (Appendix I).

Due to these discrepancies, the actual survey data were used to estimate cell volume changes for the sediment budget in this study. While it is recognised that this approach does not account for sediment transfers on the lower shoreface, the close consistency between the short-term (Section 7.2) and medium-term budgets (Section 7.1) indicates that this approach is suitable to quantify sediment transfers on the beach-face. This is considered acceptable given the nature of gravel transport on beaches. Results from tracer experiments indicate that gravel transport is most rapid on the upper beach near the high water mark (Bray 1997) and the morphological expression of slugs of gravel is on the upper beach, with remnants of slugs left at different beach levels according to the severity and sequence of storms (Brunsden 1999).

Errors due to extrapolation are more problematic on sand beaches as sand volume changes on the lower shoreface are important (Deruig and Louisse 1991). One of the failings of many sediment budget studies is the lack of any formal error analysis concerning the extrapolation of profiles to a common closure depth. This is a key area for further research given the vast beach profile data sets currently available for many coastlines.

8. CONCLUSIONS

This thesis has quantified the transfer of gravel-sized sediment from one river to the coast and its subsequent redistribution within the coastal zone. In doing so, the nature of sediment exchanges between the fluvial, deltaic and coastal systems have been established (Chapters 5, 6 and 7) and conclusions and implications for further research highlighted.

The main conclusions may be summarised as follows:

1. Bedload transport rates in the wandering gravel-bed reach of the lower River Spey were quantified using a morphological approach (cf. Ashmore and Church 1999). Sediment transport rates decreased during the study from $41\,000 \pm 6\,000 \text{ m}^3 \text{ a}^{-1}$ (1993-1994) to $6\,000 \pm 4\,000 \text{ m}^3 \text{ a}^{-1}$ (1997-1999). Although large floods generally caused the greatest morphological change at a reach scale (and hence the most transport), flow conditions were only one of the controls on transport rates. Access to available sediment was a critical factor influencing enhanced bedload transport rates and the antecedent condition of the reach prior to increased flows was important. As a result, transport rates vary dramatically for a given flow condition according to the availability of sediment for transport (for example the 3-year flood event that occurred in July 1997 had less effect on transport rates than previous floods of similar magnitude). In the Spey wandering gravel-bed channel, sediment is mobilised according to the recent local history of erosion, sediment deposition and channel adjustment and not just the magnitude of the flood. This conforms with work elsewhere (cf. McLean and Church 1999).
2. The nature of the fluvial supply results in episodic, or pulsed, delivery of sediment to the river mouth. Subsequent delivery to the coast is variable depending on the operation of the delta. A gravel spit complex extended westwards across the river mouth at a mean rate of 150 m a^{-1} between 1997 and 1999, and resulted in the temporary storage of sediment within the extending spit and in the lee of the spit, manifest as a series of low recurving gravel ridges welded onto the lower foreshore. This depositional lobe migrated downdrift, preceded by an erosional bight which also migrated downdrift at a similar rate. Interaction at the river mouth therefore appears to result in zones of erosion/accretion which migrate downdrift and can cause significant variations in foreshore morphology and longshore transport rates (cf. Kirk 1991; Hicks et al. 1999).
3. Beach volume changes estimated from successive beach profile surveys show that zones of erosion and accretion are spatially and temporally variable along the 16km coastline of Spey Bay. Spatial and temporal variability in beach volumes is caused by the passage

of pulses (slugs) of sediment which move alongshore in response to variations in sediment supply (e.g. episodic delivery of fluvial sediment, river mouth processes, beach feeding and storm events). For example, the storm event of March 1998 resulted in a sediment slug with a wavelength of ca. 2-3km and involved gravel volumes of ca. $10 \text{ m}^3 \text{ m}^{-1}$. As a result it appears that gravel clasts, although moving alongshore as individual grains, move alongshore at similar rates and so appear as a coherent mass (or slug) that has moved by a given distance depending on the magnitude of the storm. At Spey Bay the dominant direction of transport is to the west. Slugs reside on the upper beachface and are morphologically expressed as ridges, accretionary cusps and overwash lobes at different elevations. The residence time of a given slug appears to depend on its initial elevation and/or the frequency of storms of the required magnitude for movement. The response of a given stretch of beach to a particular storm is therefore variable alongshore depending on the antecedent beach condition and thus the Spey Bay profiles display something of the Markovian characteristics found on mixed-sediment beaches elsewhere (cf. Sonu and James 1973). Beach sediment is mobilised according to the recent local history of erosion and deposition and not just the magnitude of the storm. This parallels findings in wandering gravel-bed rivers (conclusion 1).

4. Longshore gravel transport rates were quantified by applying the principle of sediment continuity to estimates of beach cell volume changes. Mean estimated transport rates at the downdrift boundary were $17\,000 \text{ m}^3 \text{ a}^{-1}$ (but short-term variation between $126\,000 \pm 36\,000 \text{ m}^3 \text{ a}^{-1}$ to $-72\,000 \pm 69\,000 \text{ m}^3 \text{ a}^{-1}$ occurred). However, field evidence also indicates that gravel transport rates are negligible at this boundary as this marks the downdrift transition from a gravel to a sand beach. It is likely that onshore-offshore sediment exchanges represent the shortfall, together with much lesser amounts of counter-drift gravels.
5. The short-term (3 year) and medium-term (100 year) sediment budgets for Spey Bay both showed a net loss of sediment, indicating a system that has been erosional for at least the last century. Comparison of the short-term budget, which showed losses of ca. $16\,300 \text{ m}^3 \text{ a}^{-1}$ and $3\,600 \text{ m}^3 \text{ a}^{-1}$ of gravel and sand, respectively, with the medium-term budget, with gravel and sand losses of ca. $10\,300 \text{ m}^3 \text{ a}^{-1}$ and $4\,500 \text{ m}^3 \text{ a}^{-1}$, respectively, show good agreement. This indicates internal consistency in the data and a robust approach. The general trends of beach erosion and accretion along the coast of Spey Bay are entirely supported by the medium-term budget and, with the exception of beach accretion relating to the westward movement of the coarse-clastic beach front, the entire

coastline is erosional. The short-term budget similarly corroborates this trend, although several short sections of the coast recorded short-lived accretion within the overall erosional trend. These may relate to the passage of sediment slugs (conclusion 3). The long-term (10 000 year) budget indicated that the Spey Bay system has moved from one of net accretion during the mid-Holocene to one of net erosion and sediment deficit in the late Holocene. This decline in sediment supply has led to a sediment cell dominated by the internal re-organisation processes of erosion and deposition (cf. Hansom 2000) as deposition is predominantly fuelled by the updrift erosion of pre-existing beach deposits.

6. A conceptual model of gravel sediment transfers from river to beaches is proposed, which highlights the episodic nature of gravel transport. Fluvial sediment supply is variable and depends not only on flood magnitude and frequency but also on the availability of sediment for mobilisation (conclusion 1). Interactions at the river mouth result in cycles of erosion/accretion along the downdrift coast (conclusion 2) and gravel is moved along the coast as a series of slugs relating to pulsed variations in updrift sediment supply and storms (conclusion 3). This model has important implications for the management of gravel beaches as erosional (or “problem”) sections of shoreline are neither temporally nor spatially static. For example, a stretch of beach which is undergoing erosion and loss of sediment at time 1 may undergo accretion at time 2 as a sediment slug migrates downdrift. Protecting an erosional section with a structure of some sort will result in adjustment of the natural balance of the beach system, leading to a progressive cascade of problems downdrift.

Several conclusions relate more specifically to the methodology used in this study. These are summarised below:

7. Cross-sections spaced at 325m (ca. twice the active width) on the lower River Spey provide inaccurate estimates of sub-reach storage changes over a three year period using the method of Martin and Church (1995). A denser survey network is required on rivers with similar scale and characteristics to the Spey.
8. In order to quantify the volume of gravel bars ($D_{50} = 0.04$) to within $\pm 2.5\%$ of the true volume a survey point spacing of 6m is recommended (or a spacing/ $\sqrt{\text{area}}$ ratio of less than 0.15). Percentage errors increase substantially when this ratio is exceeded.
9. Previous studies that use successive surveys of beach profiles to estimate volume changes in the cells they delimit give no indication of the errors associated with varying

profile spacing (e.g. Comber 1993; Hicks et al. 1999). Detailed analysis of closely spaced beach profiles in this study indicate that errors of up to $\pm 118\%$ in the estimated cell volume change can occur at a longshore spacing of 500m, although the percentage error declines with the magnitude of the change. Large volumetric changes are likely to be associated with smaller percentage errors.

10. From conclusions 7 and 9, error analysis must always be reported in sediment budget studies.

This study has highlighted numerous possibilities for further analysis and research. These may be summarised as follows:

1. The proposed model describing gravel sediment transfers from rivers to beaches provides a conceptual basis for development of a predictive model whereby the magnitude and wavelength of sediment slugs are predicted for different magnitudes of river floods and coastal storms. Such a model must also account for the antecedent condition of the river and beach sediment prior to the flood/storm in order to infer its mobilisation potential. There may thus be a role for Markovian models in prediction of this nature. Further research on smaller scale fluvial-coastal gravel systems, small sections of river or coast, or physical models, could advance such a model.
2. This work has advanced the current knowledge of the dynamics of “small” river mouths (cf. Zenkovitch 1967; Kirk 1991). However there is scope to further investigate the nature of the cycles of erosion/accretion generated along the downdrift coast in response to river mouth processes. A much denser temporal and spatial survey of the river mouth and downdrift coast than was achievable in this study would allow the details of this process to be interpreted. Recent developments in photogrammetry (Lane et al. 1993; Chandler 1999) and Global Positioning Systems (GPS) (Acharya and Chaturvedi 1997) may allow speedier acquisition of such data-sets at river mouths.
3. Current theoretical approaches to gravel sediment transport fail to account for complexities related to the morphological and structural characteristics of the sediment pile. This research has shown that river and beach sediment is mobilised according to the recent local history of erosion and sediment deposition and not only the magnitude of flood/storm. Further research is required to fully assess the role of these other factors that drive sediment exchanges.
4. This research has highlighted several methodological issues related to sediment budgeting which require further research. Further research is required to assess the

density of cross-sections and beach profiles necessary to estimate volume changes to within a given level of accuracy at different scales. In addition, the “closure depth” of gravel remains inadequately defined and errors in the extrapolation of profiles out to arbitrary depths may be a consistent source of error in sediment budget studies. Given the vast beach profile data sets currently available for many coastlines this is a key area for further research.

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APPENDIX A: DATA AVAILABILITY

River	D	J	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M
	9	9	5	6											9	8			
geomorphological mapping	1													2					
cross section surveys	1	1										2							
bar surveys																			
survey point spacing test																			1
Delta																			
sub-aerial survey																	1		2
sub-aqueous survey																			
Beach																			
geomorphological mapping	1																		
beach profile surveys					1	1					2						3		
profile spacing tests																			
offshore extension of profiles																			
River Floods																			
Coastal Storms																			

¹ Wetted channel not surveyed
² Incomplete due to the early July flood
³ Due to access problems a small section of survey 1 delayed until September 1997
⁴ Not as extensive as survey 1, only the areas which had visibly changed were surveyed
⁵ Survey of 8 profiles at 100m spacing (locations -2.3 to -1.6)

APPENDIX A: DATA AVAILABILITY (CONTINUED)

River	J 9 8	J	A	S	O	N	D	J 9 9	F	M	A	M	J	J	A 9 9
geomorphological mapping											3				
cross section surveys												3			3
bar surveys	3														
survey point spacing test															
Delta															
sub-aerial survey															
sub-aqueous survey			1 ⁶												2
Beach															
geomorphological mapping															
beach profile surveys	7	7		8					9						
profile spacing tests				2 ⁷											
offshore extension of profiles			1 ⁸												2
River Floods															
Coastal Storms															

⁶ Swell conditions

⁷ Survey of 10 profiles at 100m spacing (locations -12 to -11)

⁸ Incomplete (profiles +4 to +0 and -12 to -8.5 only). Swell conditions during this survey limit its accuracy

APPENDIX B : Sub-reach storage changes (δS) from planimetric channel changes and reach-scale sediment budget for the lower River Spey

Erosion and deposition are given as bulk volumes. Sub-reach storage changes (δS) and volumetric transport rates into and out of each sub-reach (Q_i and Q_o) are in mineral volumes (i.e. corrected for porosity).

July 1993 - July 1994

Reach	erosion (m ³)	deposition (m ³)	δS (m ³)	Q_i (m ³)	Q_o (m ³)
1	13 590	8 751	-3 387	3 276	6 663
2	73	1 231	+810	6 663	5 853
3	11 298	9 133	-1 516	5 853	7 369
4	63	1 183	+784	7 369	6 585
5	16 845	10 819	-4 218	6 585	10 803
6	22 878	17 716	-3 613	10 803	14 417
7	33 522	23 006	-7 361	14 417	21 778
8	42 550	26 163	-11 471	21 778	33 249
9	29 575	14 241	-10 734	33 249	43 983
10	25 123	16 889	-5 764	43 983	49 747
11	13 921	26 045	+8 486	49 747	41 261

volumetric
transport rate out
of downstream
sub-reach
(m³ a⁻¹)

41 261 ± 6 360
m³ a⁻¹

July 1994 - December 1995

Reach	erosion (m ³)	deposition (m ³)	δS (m ³)	Q_i (m ³)	Q_o (m ³)
1	0	0	0	3 173	3 173
2	619	0	-433	3 173	3 606
3	8 770	3 179	-3 914	3 606	7 519
4	2 617	1 773	-591	7 519	8 110
5	5 094	0	-3 566	8 110	11 676
6	20 112	4 638	-10 832	11 676	22 508
7	21 620	11 971	-6 754	22 508	29 262
8	10 835	2 454	-5 867	29 262	35 129
9	10 837	12 797	+1 372	35 129	33 757
10	44 630	34 570	-7 042	33 757	40 799
11	3 692	3 493	-139	40 799	40 938

29 773 ± 6 708
m³ a⁻¹

December 1995 - December 1997

Reach	erosion (m ³)	deposition (m ³)	δS (m ³)	Qi (m ³)	Qo (m ³)
1	13 246	8 134	-3 578	14 000	17 578
2	5	12 909	+9 033	17 578	8 545
3	15 513	16 431	+642	8 545	7 903
4	1 499	3 790	+1 604	7 903	6 299
5	6 638	0	-4 647	6 299	10 946
6	24 033	5 133	-13 230	10 946	24 176
7	13 571	8 694	-3 414	24 176	27 589
8	4 841	5 929	+762	27 589	26 828
9	7 253	10 523	+2 289	26 828	24 539
10	27 114	23 211	-2 732	24 539	27 271
11	23 359	6 451	-11 836	27 271	39 107

volumetric
transport rate out
of downstream
sub-reach
(m³ a⁻¹)

19 553 ± 3 687
m³ a⁻¹

December 1997 - April 1999

Reach	erosion (m ³)	deposition (m ³)	δS (m ³)	Qi (m ³)	Qo (m ³)
1	7 913	6 543	-959	2 686	3 645
2	1 331	301	-721	3 645	4 366
3	6 890	1 374	-3 861	4 366	8 228
4	3 130	6 248	+2 183	8 228	6 045
5	10 048	12 710	+1 863	6 045	4 182
6	11 455	13 873	+1 692	4 182	2 490
7	7 484	1 675	-4 067	2 490	6 557
8	3 478	0	-2 435	6 557	8 991
9	8 537	4 866	-2 570	8 991	11 561
10	11 809	20 723	+6 240	11 561	5 321
11	16 520	12 117	-3 083	5 321	8 404

5 932 ± 4 107
m³ a⁻¹

Sub-reach storage changes (δS) from cross-section surveys
(October 1997 to May 1999)

Reach	δS (m ³)	Qi (m ³)	Qo (m ³)
1	-52	2832	2885
2	+2 149	2885	736
3	+4 015	736	-3279
4	+2 895	-3279	-6175
5	+3 723	-6175	-9898
6	-747	-9898	-9151
7	-371	-9151	-8780
8	+834	-8780	-9614
9	-1 469	-9614	-8145
10	-20 464	-8145	12318

7 820 m³ a⁻¹

APPENDIX C : Summary River Surface Particle Size Analysis

Profile location samples (100 random surface clasts)

sections	dist. from mouth/ m	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
1	5764	-5.49	44.94	0.86	-0.05	-5.54	46.53	-4.62	24.59	-6.30	78.79	1.79
2	5334	-4.51	22.78	0.70	0.05	-4.50	22.63	-3.78	13.74	-5.25	38.05	1.66
3	5009	-5.50	45.25	0.95	-0.03	-5.51	45.57	-4.54	23.26	-6.46	88.03	1.95
4	4759	-5.04	32.90	0.88	-0.02	-5.07	33.59	-4.13	17.51	-5.94	61.39	1.87
5	4553	-4.87	29.24	0.78	0.11	-4.83	28.44	-4.12	17.39	-5.67	50.91	1.71
6	4243	-5.08	33.82	0.82	0.09	-5.03	32.67	-4.24	18.90	-5.98	63.12	1.83
7	3860	-5.71	52.35	0.68	0.01	-5.69	51.63	-5.06	33.36	-6.38	83.29	1.58
8	3653	-5.87	58.49	0.79	-0.01	-5.86	58.08	-5.06	33.36	-6.69	103.25	1.76
9	3354	-5.82	56.49	0.73	0.02	-5.80	55.72	-5.09	34.06	-6.57	95.01	1.67
10	2703	-5.60	48.50	0.82	-0.01	-5.63	49.52	-4.71	26.17	-6.45	87.43	1.83
11	2288	-5.91	60.13	0.80	-0.07	-5.94	61.39	-5.10	34.30	-6.69	103.25	1.74
12	1990	-5.23	37.53	0.82	-0.12	-5.30	39.40	-4.34	20.25	-6.06	66.72	1.82
13	1363	-4.90	29.86	0.82	0.08	-4.83	28.44	-4.09	17.03	-5.79	55.33	1.80
14	903	-5.32	39.95	0.73	-0.28	-5.47	44.32	-4.50	22.63	-5.97	62.68	1.66
15	400	-4.91	30.06	0.88	0.14	-4.83	28.44	-4.09	17.03	-5.79	55.33	1.80

Bar-head samples (400 surface clasts sampled from a 0.5 x 0.5m area)

sections	dist. from mouth/ m	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
10 (left bar)	2703	-5.01	32.22	0.69	-0.05	-5.00	32.00	-4.29	19.56	-5.73	53.08	1.65
11 (right bar)	2288	-5.41	42.52	0.84	-0.05	-5.35	40.79	-4.59	24.08	-6.30	78.79	1.81
12 (left bar)	1763	-4.96	31.12	0.64	-0.05	-4.93	30.48	-4.34	20.25	-5.62	49.18	1.56
13.5 (left bar)	1163	-5.25	38.05	0.68	-0.05	-5.24	37.79	-4.59	24.08	-5.93	60.97	1.59
14.5 (right bar)	800	-5.33	40.22	0.68	-0.05	-5.29	39.12	-4.66	25.28	-6.04	65.80	1.61
15 (left bar)	400	-4.69	25.81	0.81	-0.05	-4.65	25.11	-3.92	15.14	-5.49	44.94	1.72

Sand Samples (<2mm)

sections	dist. from mouth/ m	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
6	4243	1.33	0.40	0.80	-0.05	1.36	0.39	2.03	0.24	0.61	0.66	1.64
9	3354	0.75	0.59	0.81	0.04	0.76	0.59	1.48	0.36	0.01	0.99	1.66
12	1990	1.75	0.30	0.50	-0.07	1.74	0.30	2.25	0.21	1.26	0.42	1.41
13	1363	0.52	0.70	1.32	0.23	0.83	0.56	1.84	0.28	-1.11	2.16	2.78
14	903	1.55	0.34	0.51	0.04	1.57	0.34	1.98	0.25	1.09	0.47	1.36

APPENDIX D : Bulk Particle Size Analysis (River, Delta and Beach)

Summary River Bulk Particle Size Analysis

Surface	dist. from mouth (m)	total mass (kg)	mass < 2mm (kg)	% sand	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
Essil	2500	1016.6	172.62	16.98	-3.84	14.32	2.75	-0.29	-4.39	20.97	-0.77	1.71	-6.36	82.14	6.94
Bar 8	1500	817.22	115.79	14.17	-4.48	22.32	2.49	-0.55	-5.36	41.07	-1.45	2.73	-6.63	99.04	6.02
Bar_b	1050	1069.42	72.82	6.81	-5.21	37.01	1.88	-0.42	-5.51	45.57	-3.44	10.85	-6.68	102.54	3.06
Bar 13	350	616.24	143.11	23.22	-3.37	10.34	2.85	-0.35	-4.15	17.75	0.21	0.86	-6.17	72.00	9.14
Sub-surface															
Essil	2500	545.12	121.43	22.28	-3.06	8.34	2.57	-0.43	-3.92	15.14	-0.17	1.13	-5.42	42.81	6.93
Bar 8	1500	732.06	89.84	12.27	-4.43	21.56	2.38	-0.46	-5.11	34.54	-1.69	3.23	-6.48	89.26	5.26
Bar_b	1050	338.55	52.52	15.51	-4.04	16.45	2.45	-0.45	-4.75	26.91	-1.14	2.20	-6.23	75.06	5.83
Bar 13	350	318.07	89.36	28.10	-2.98	7.89	2.79	-0.38	-3.87	14.62	0.63	0.65	-5.69	51.63	8.96

Summary Delta Bulk Particle Size Analysis

Surface	total mass (kg)	mass < 2mm (kg)	% sand	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
Tugnet	505.92	43.73	8.64	-5.47	44.32	1.73	-0.49	-5.75	53.82	-4.03	16.34	-6.65	100.43	2.48
Kingston s/w	976.9	6.30	0.65	-5.26	38.32	1.08	-0.14	-5.32	39.95	-4.19	18.25	-6.27	77.17	2.06
Kingston l/w	422.75	0.00	0.00	-5.19	36.50	0.95	-0.08	-5.23	37.53	-4.20	18.38	-6.15	71.01	1.97
Sub-surface														
Tugnet	393.08	31.94	8.13	-4.64	24.93	1.72	-0.28	-4.76	27.10	-3.19	9.13	-5.98	63.12	2.63
Kingston s/w	444.18	3.22	0.72	-4.86	29.04	1.15	-0.28	-5.05	33.13	-3.63	12.38	-5.9	59.71	2.2
Kingston l/w	243.62	0.00	0.00	-4.79	27.67	0.95	-0.02	-4.81	28.05	-3.80	13.93	-5.75	53.82	1.97

Summary Beach Bulk Particle Size Analysis

profiles	location	total mass (kg)	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)
3.5	main storm crest	748.69	-4.06	16.68	1.96	-0.14	-4.07	16.80	-2.36	5.13	-5.76	54.19	3.25
0.5	main storm crest	225.20	-5.56	47.18	0.52	-0.02	-5.57	47.50	-5.06	33.36	-6.06	66.72	1.42
-2	HWM	250.40	-4.96	31.12	0.62	0.04	-4.95	30.91	-4.29	19.56	-5.64	49.87	1.59
-6	main storm crest	225.20	-4.96	31.12	0.48	0.24	-4.88	29.45	-4.55	23.43	-5.45	43.71	1.36
-8.5	below HWM	79.30	-4.85	28.84	0.46	0.06	-4.83	28.44	-4.40	21.11	-5.33	40.22	1.38
-8.5	main storm ridge	262.40	-5.68	51.27	0.59	-0.13	-5.72	52.71	-5.03	32.67	-6.29	78.25	1.55

APPENDIX E : Error Analysis of Fluvial Sediment Budget

Plan errors, errors in the estimation of d (the depth of mobile sediment) and errors in the application of the water level correction (WL) are given as bulk volumes (m^3). All other errors are given as mineral volumes (i.e. corrected for porosity).

July 1993 - July 1994

Reach	storage change error		WL	RMS	porosity error	Qi error	Cumulative error	error at downstream reach ($m^3 a^{-1}$)
	plan	d						
1	1 503	335	na	1 079	-169	3 276	3 453	
2	264	19	na	186	41	3 453	3 458	
3	1 472	306	na	1 054	-76	3 458	3 616	
4	271	18	na	190	39	3 616	3 621	
5	1 876	416	na	1 345	-211	3 621	3 869	
6	2 128	609	na	1 556	-181	3 869	4 174	
7	2 868	848	na	2 097	-368	4 174	4 685	
8	2 920	1 029	na	2 181	-574	4 685	5 199	
9	3 444	657	na	2 460	-537	5 199	5 777	
10	3 153	822	na	2 291	-288	5 777	6 221	
11	1 678	600	na	1 254	424	6 221	6 361	6 361 $m^3 a^{-1}$

July 1994 - December 1995

Reach	storage change error		WL	RMS	porosity error	Qi error	Cumulative error	
	plan	d						
1	0	0	na	0	0	3 173	3 173	
2	893	9	na	76	-22	3 173	3 174	
3	1 155	180	na	821	-196	3 174	3 284	
4	588	66	na	415	-30	3 284	3 310	
5	740	76	na	522	-178	3 310	3 356	
6	2 557	371	1 663	2 569	-542	3 356	4 261	
7	2 307	503	3 161	3 077	-338	4 261	5 267	
8	1 669	200	1 234	1 743	-293	5 267	5 555	
9	2 108	355	3 061	2 892	69	5 555	6 264	
10	4 803	1 187	6 856	6 565	-352	6 264	9 080	
11	1 108	108	1 855	1 620	-7	9 080	9 224	6 708 $m^3 a^{-1}$

December 1995 - December 1997

Reach	storage change error		WL	RMS	porosity error	Qi error	Cumulative error	
	plan	d						
1	2 652	321	na	1 870	-179	2 289	2 962	
2	1 029	194	na	734	452	2 962	3 084	
3	2 566	479	na	1 832	32	3 084	3 588	
4	921	79	na	648	80	3 588	3 646	
5	1 081	100	na	761	-232	3 646	3 732	
6	4 016	437	2 084	3 494	-661	3 732	5 155	
7	2 719	334	1 504	2 396	-171	5 155	5 687	
8	1 554	162	641	1 298	38	5 687	5 833	
9	2 125	267	1 342	1 926	114	5 833	6 144	
10	3 801	755	2 936	3 643	-137	6 144	7 144	
11	2 048	447	778	1 726	-592	7 144	7 374	3 687 $m^3 a^{-1}$

APPENDIX E (continued) : Error Analysis of Fluvial Sediment Budget

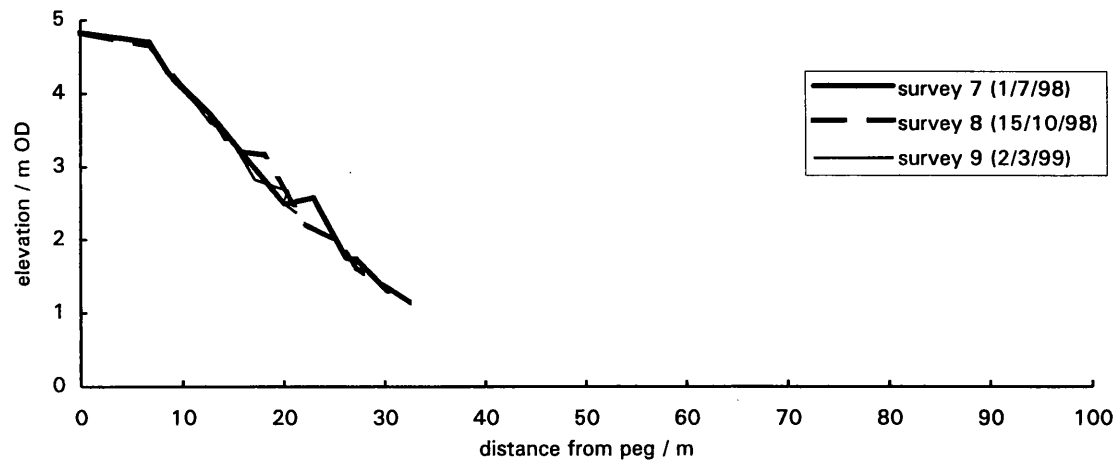
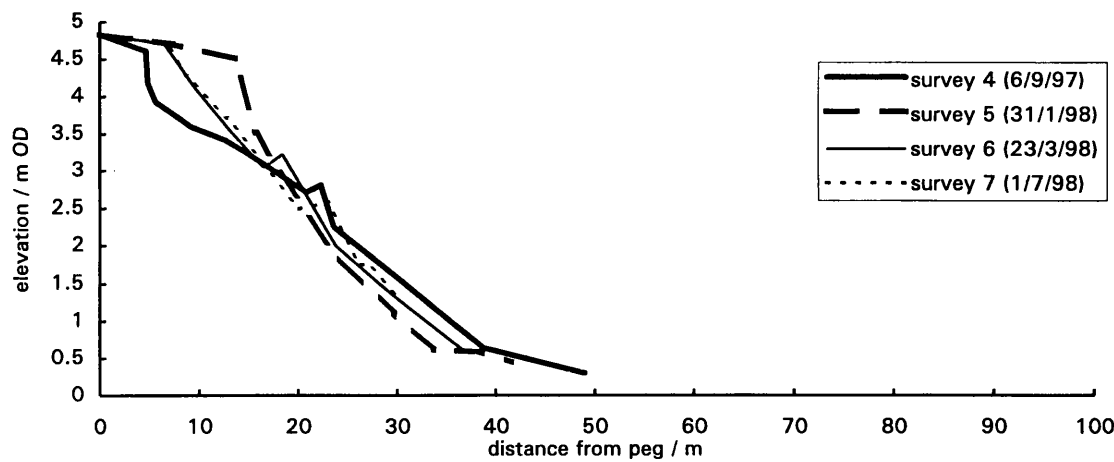
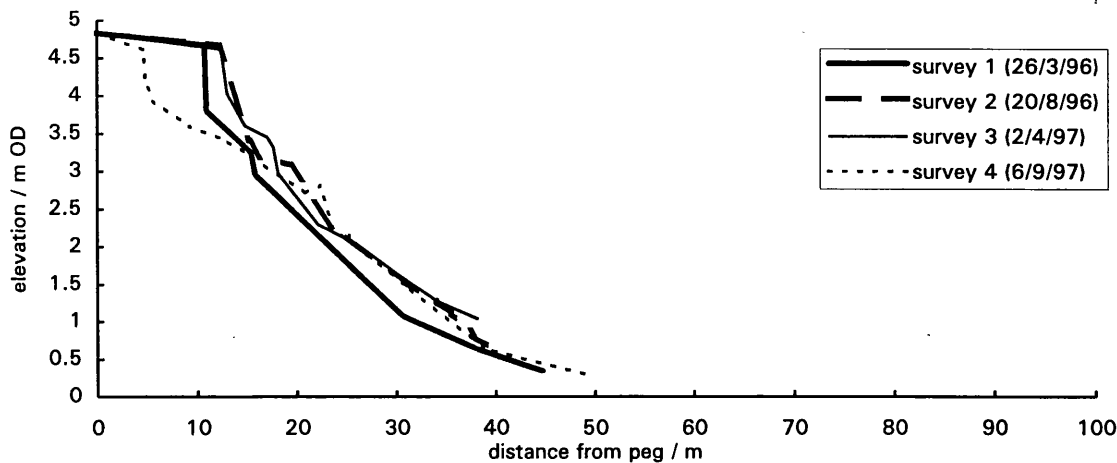
December 1997 - April 1999

Reach	storage change error		WL	RMS	porosity error	Qi error	Cumulative error
	plan	d					
1	1 933	217	na	1 362	-48	2 686	3 012
2	427	24	na	300	-36	3 012	3 027
3	1 299	124	na	914	-193	3 027	3 168
4	1 808	141	na	1 270	109	3 168	3 415
5	2 954	341	na	2 084	93	3 415	4 002
6	4 394	380	na	3 089	85	4 002	5 056
7	2 065	137	na	1 450	-203	5 056	5 264
8	660	52	na	463	-122	5 264	5 285
9	1 207	201	na	857	-128	5 285	5 356
10	2 202	488	na	1 584	312	5 356	5 594
11	2 228	430	na	1 591	-154	5 594	5 818

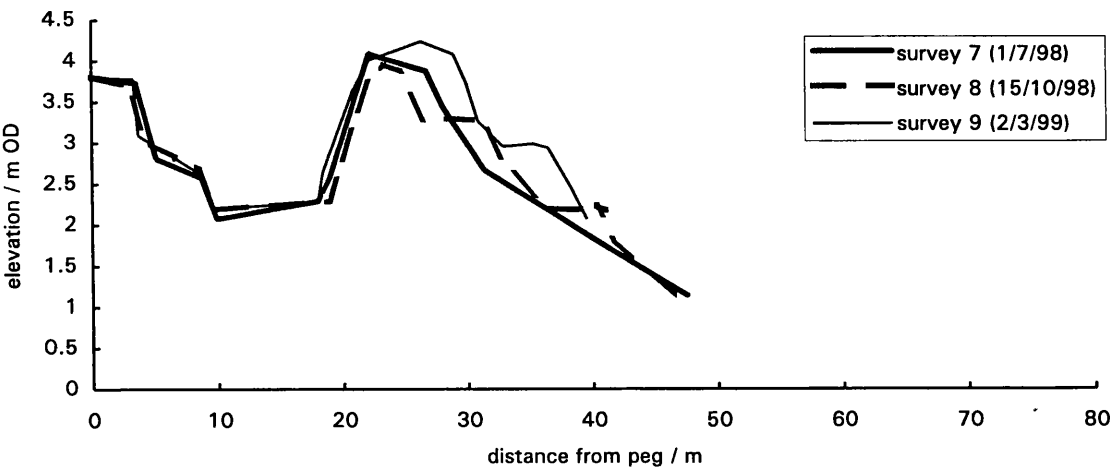
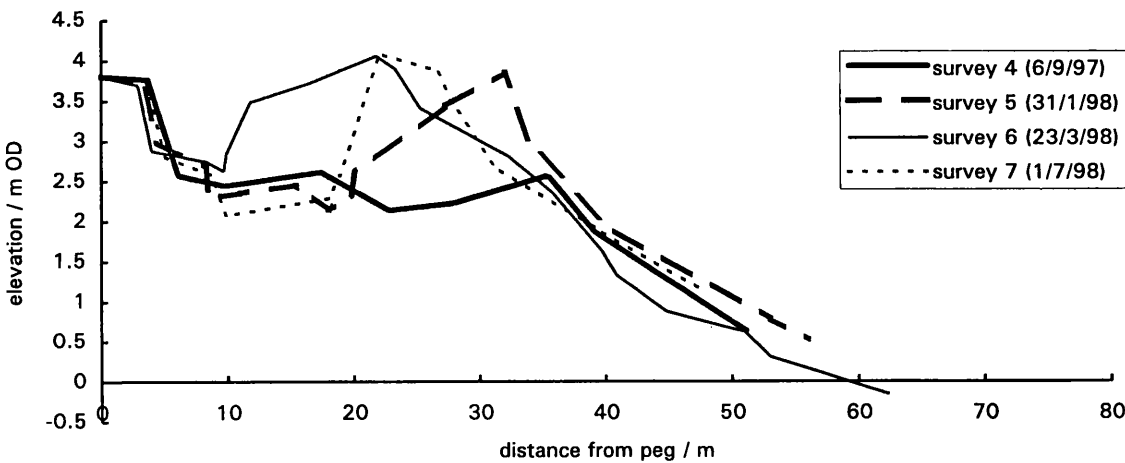
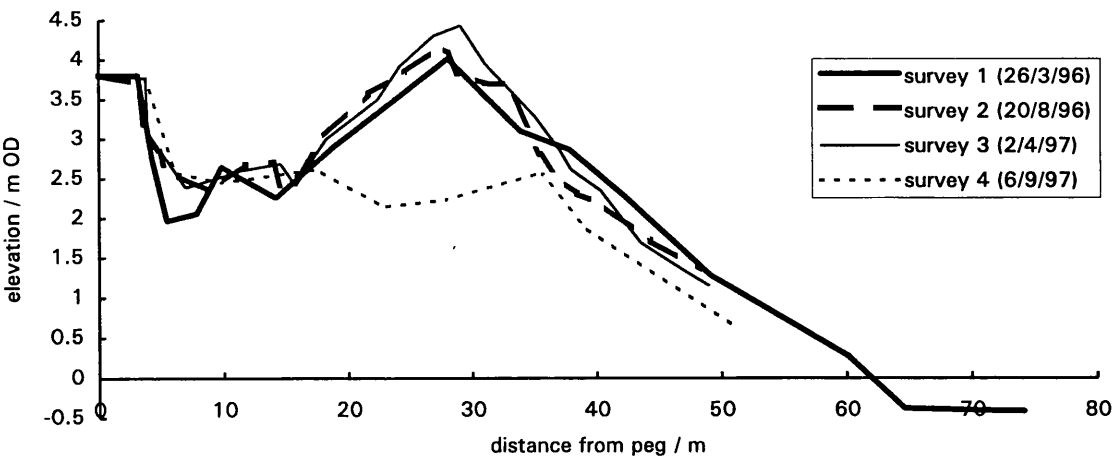
4 107 m³ a⁻¹

APPENDIX F : BEACH PROFILE CHANGE

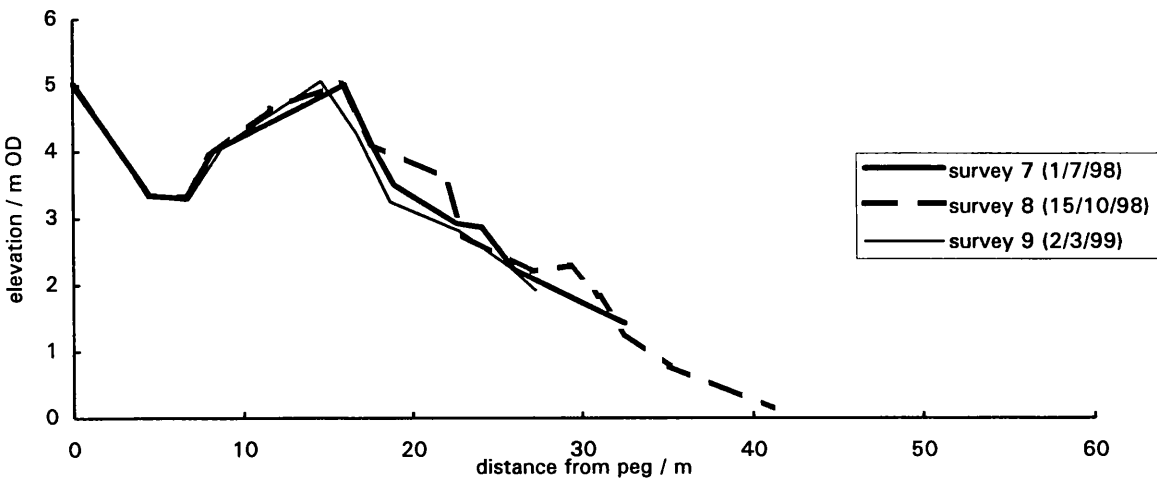
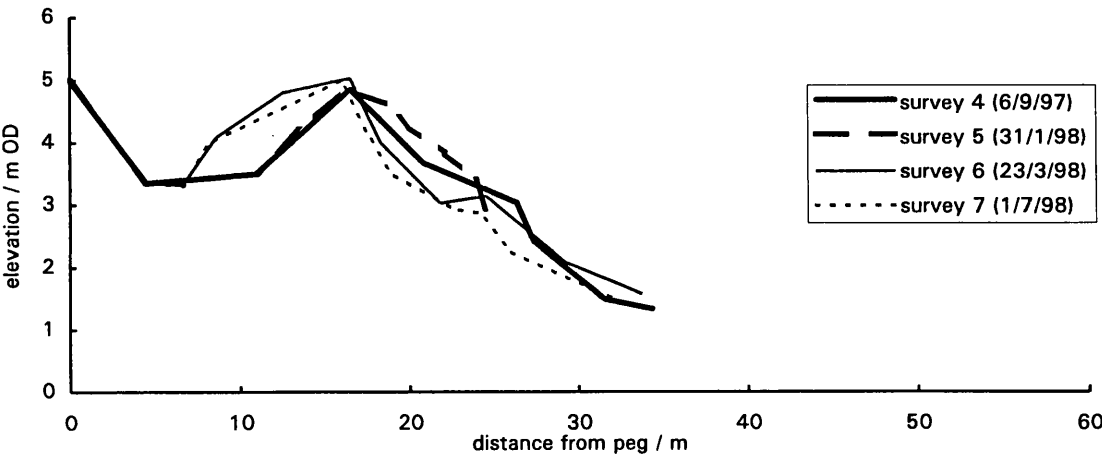
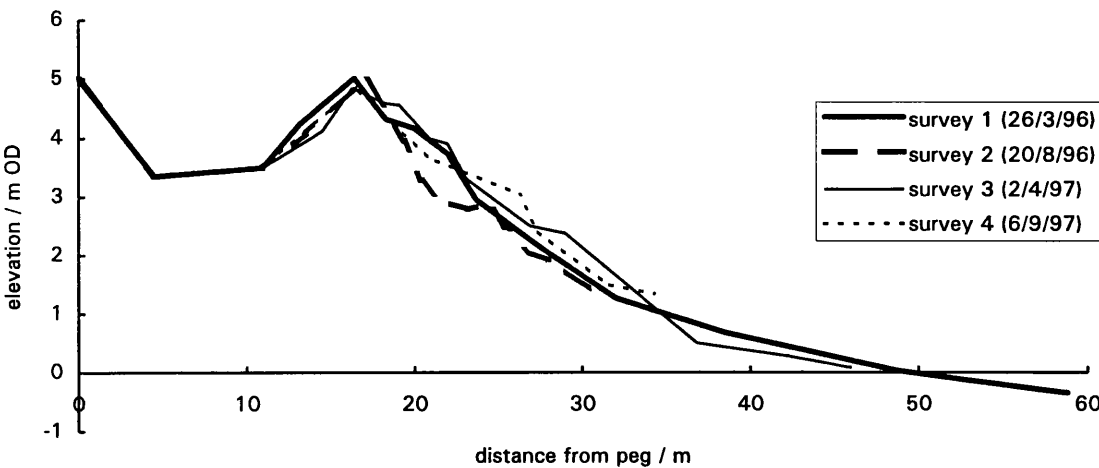
Profile +4: Morphological Change (March 1996 to March 1999)



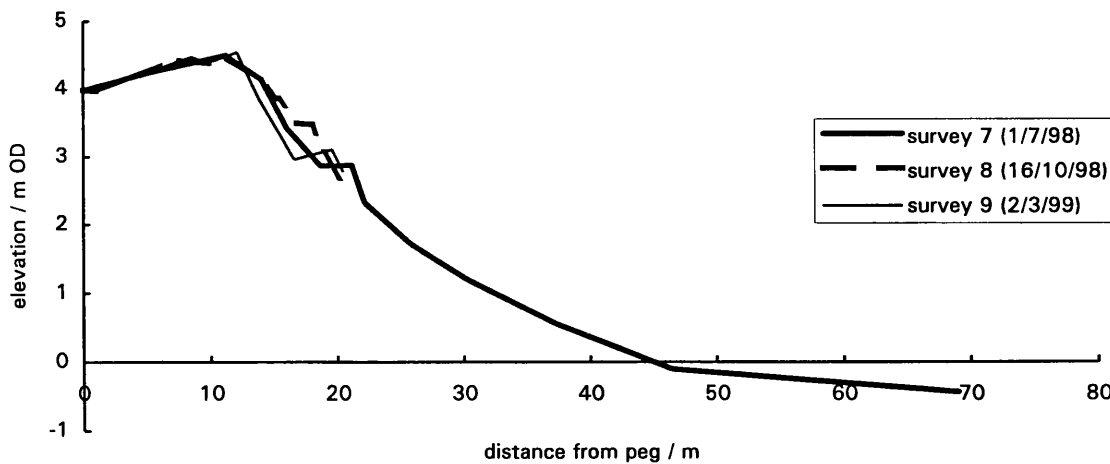
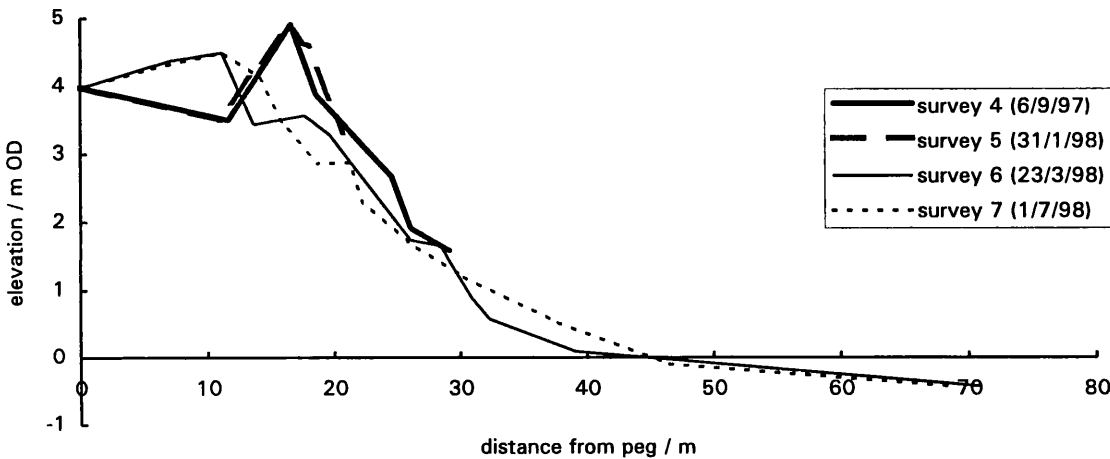
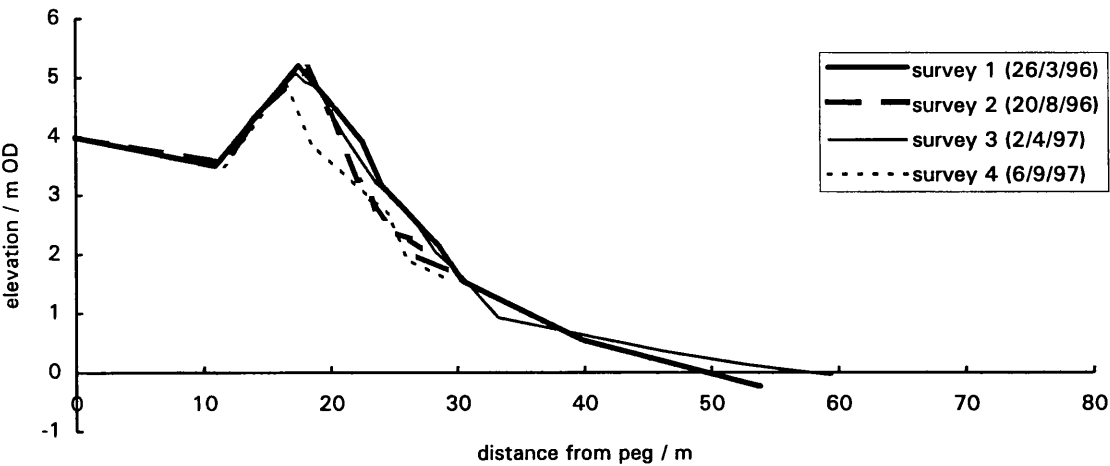
Profile +3.5 (Tynet Burn): Morphological Change (March 1996 to March 1999)



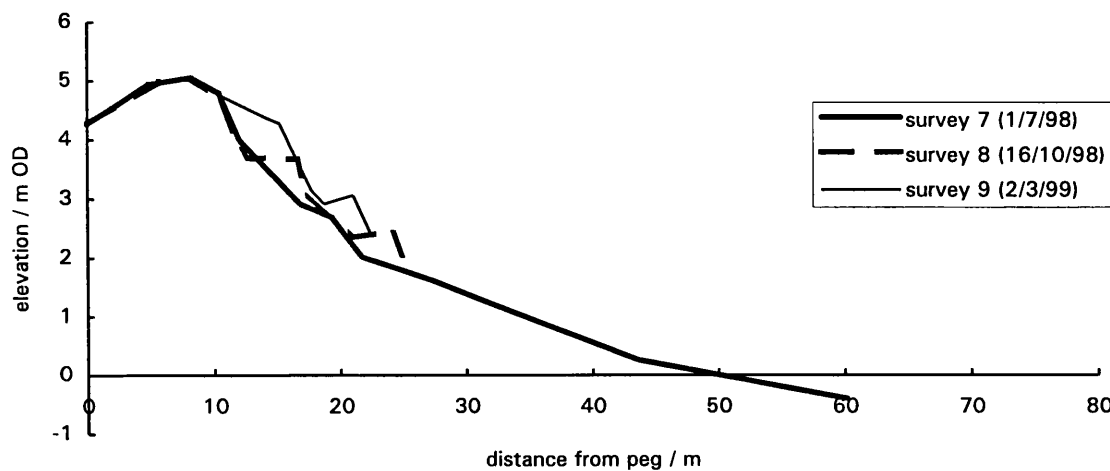
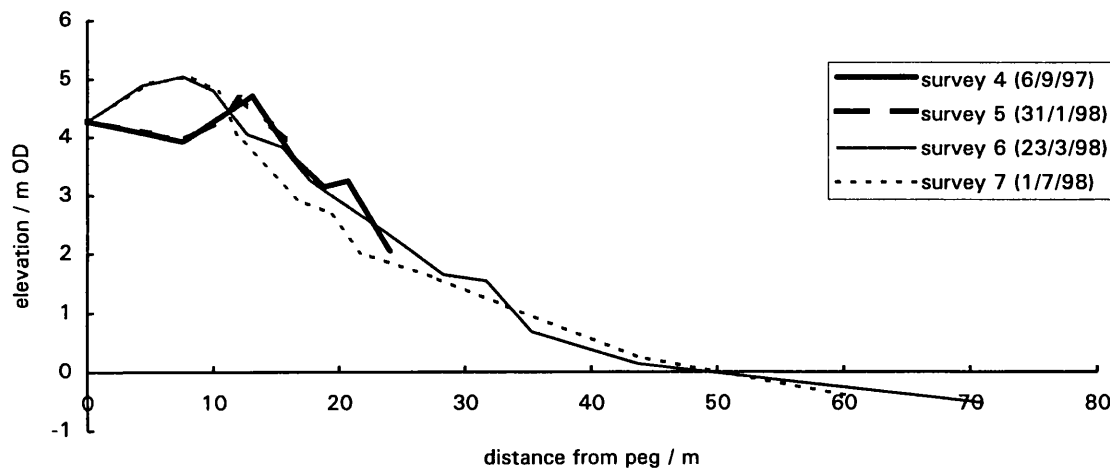
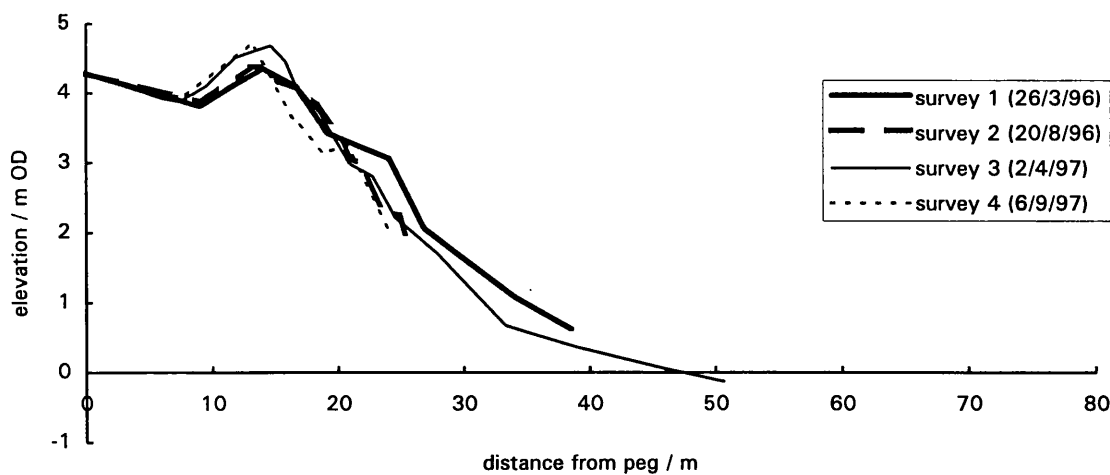
Profile +3: Morphological Change (March 1996 to March 1999)



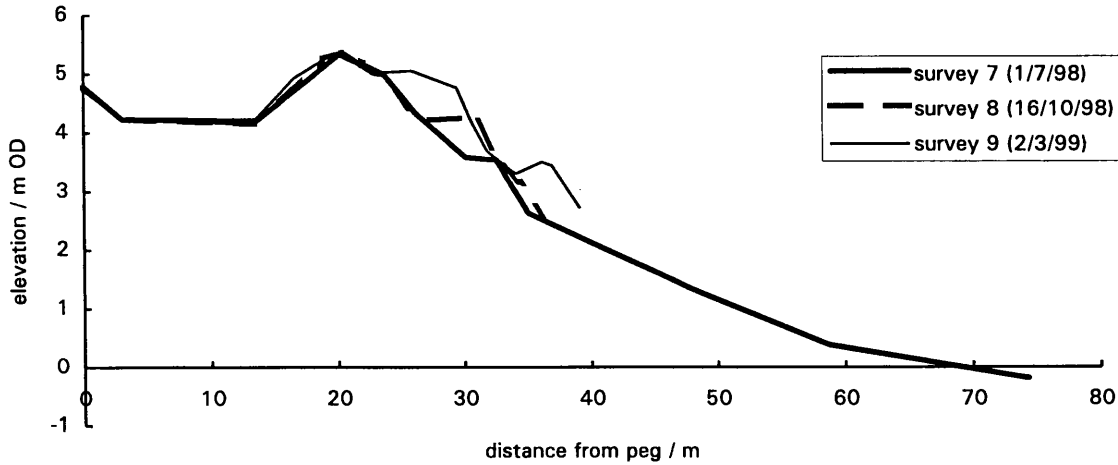
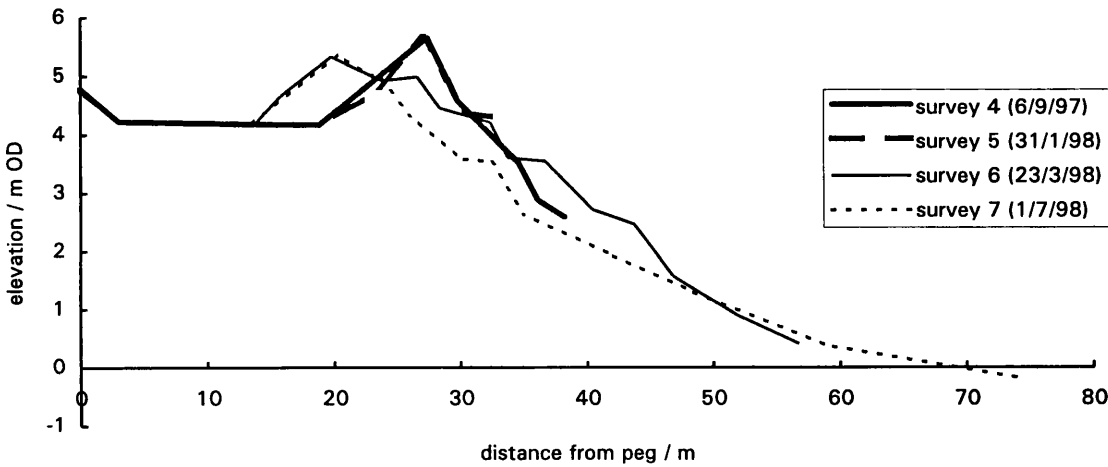
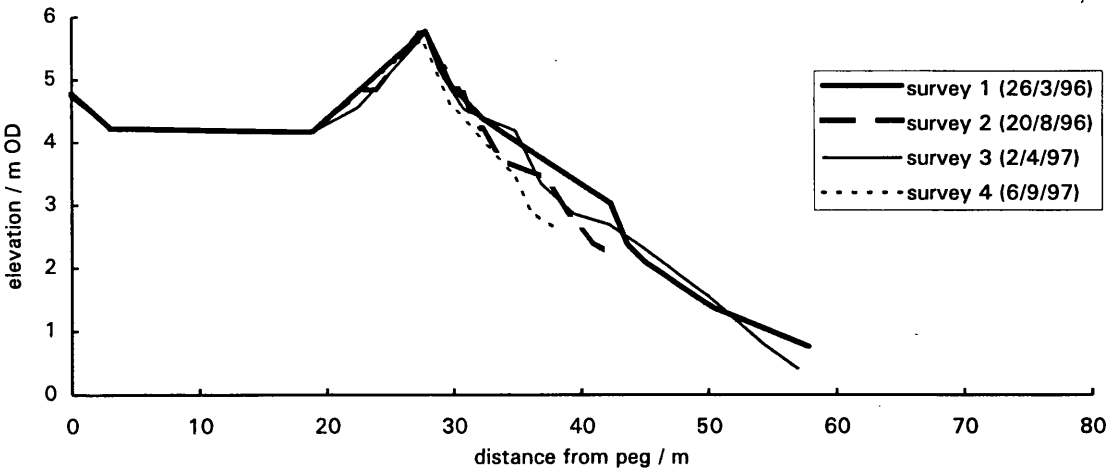
Profile +2.5: Morphological Change (March 1996 to March 1999)



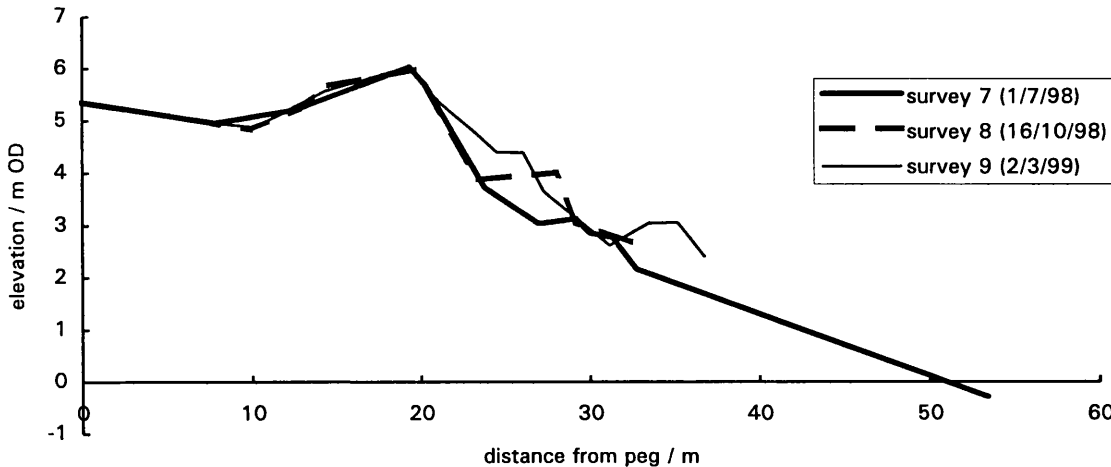
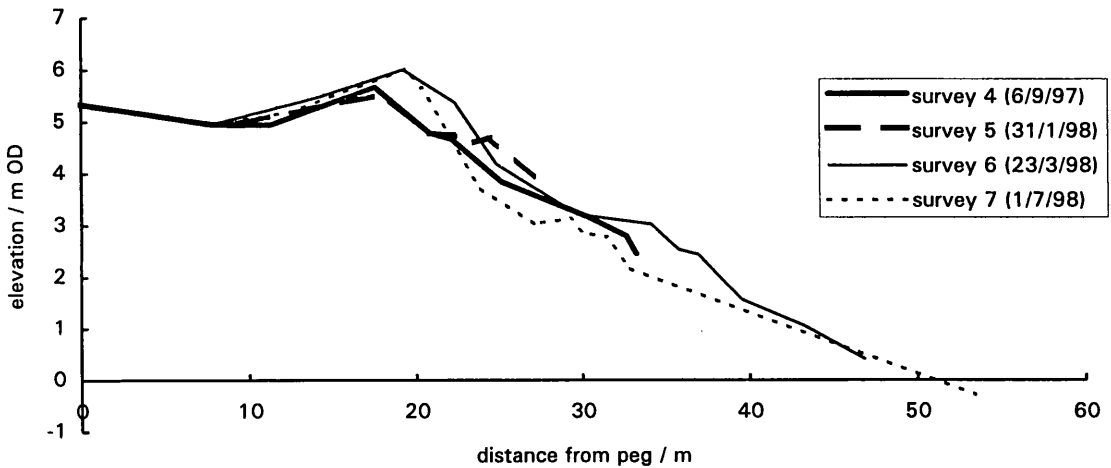
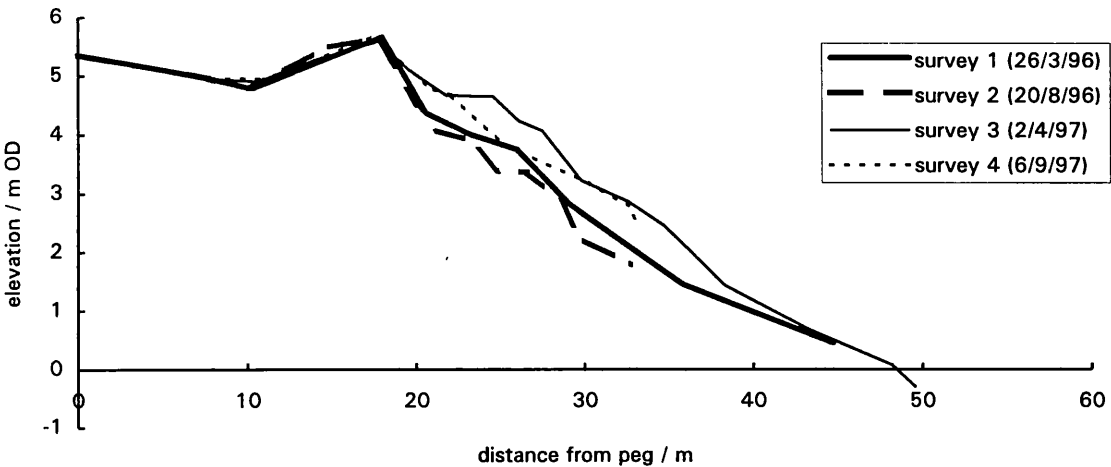
Profile +2: Morphological Change (March 1996 to March 1999)



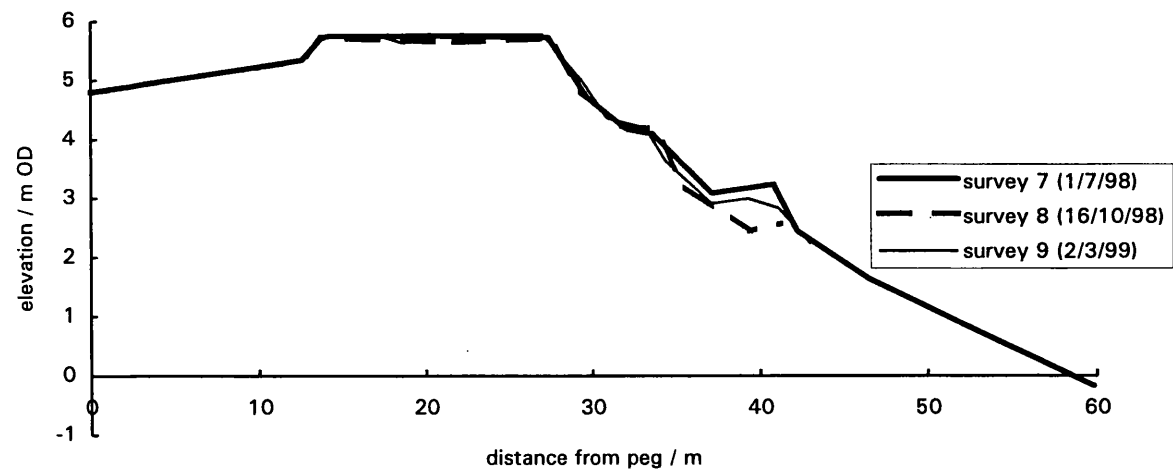
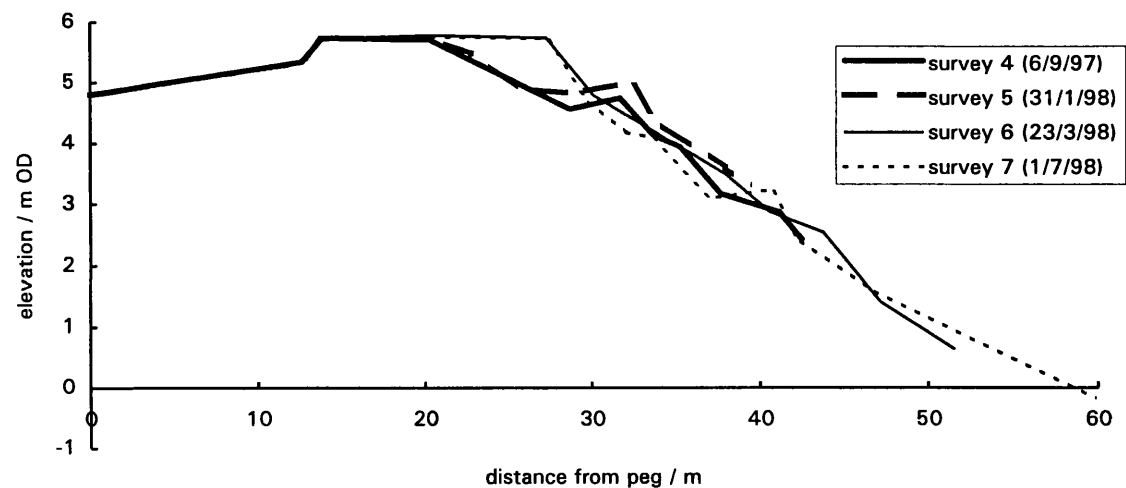
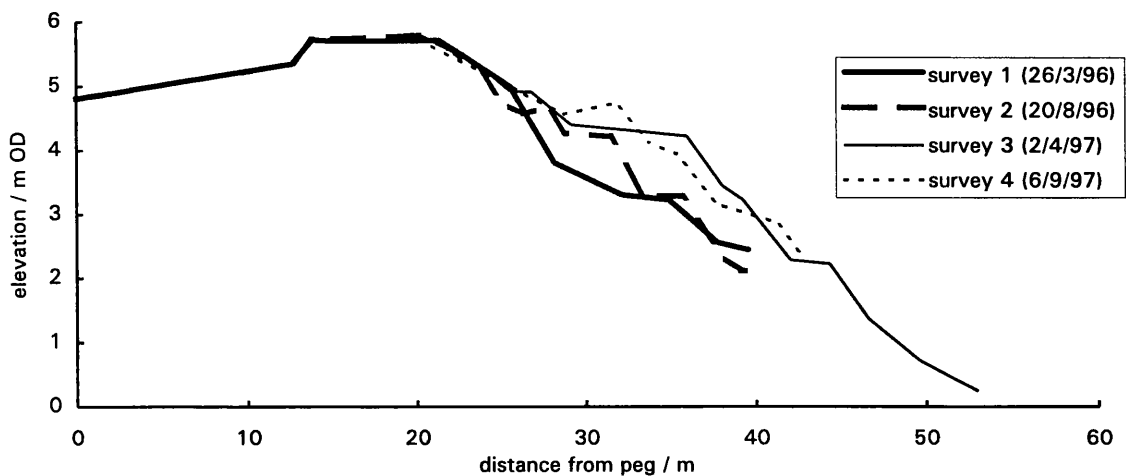
Profile +1.5: Morphological Change (March 1996 to March 1999)



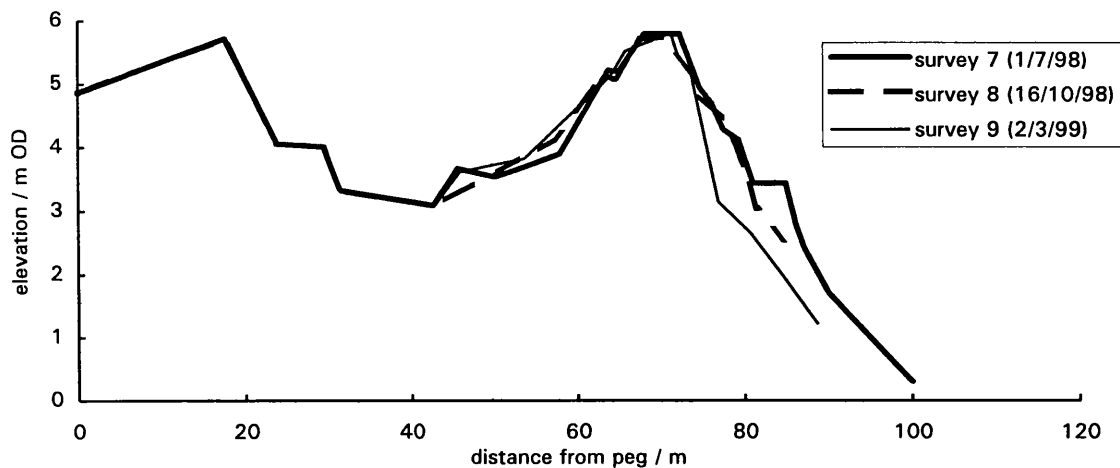
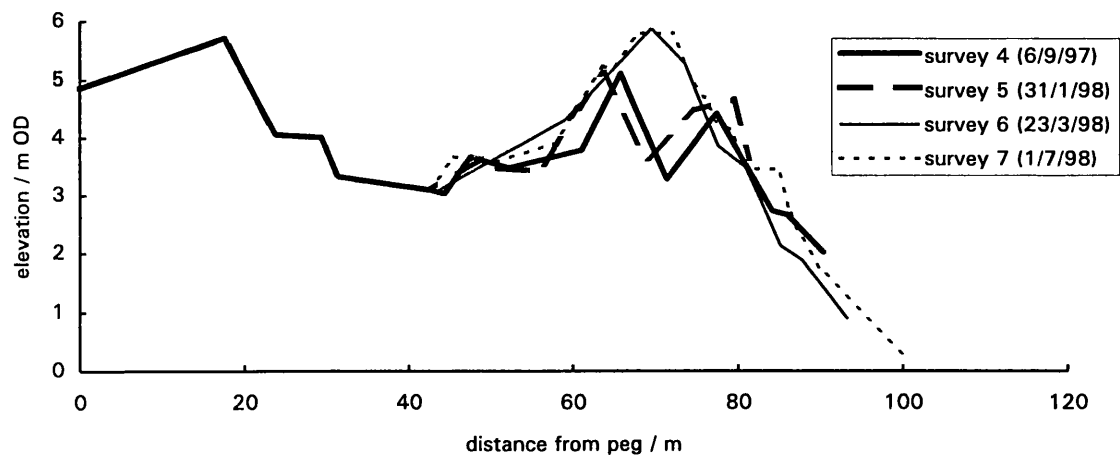
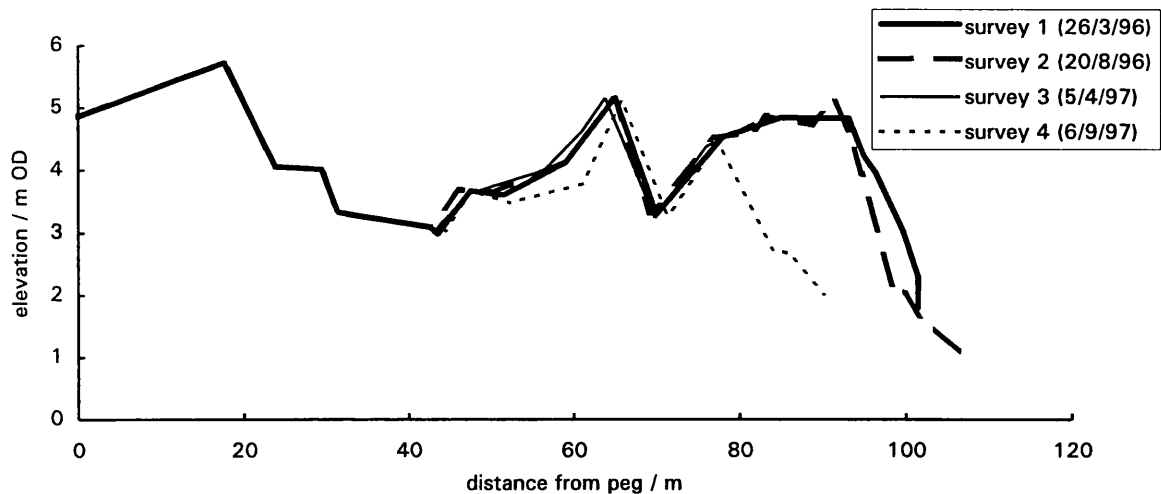
Profile +1: Morphological Change (March 1996 to March 1999)



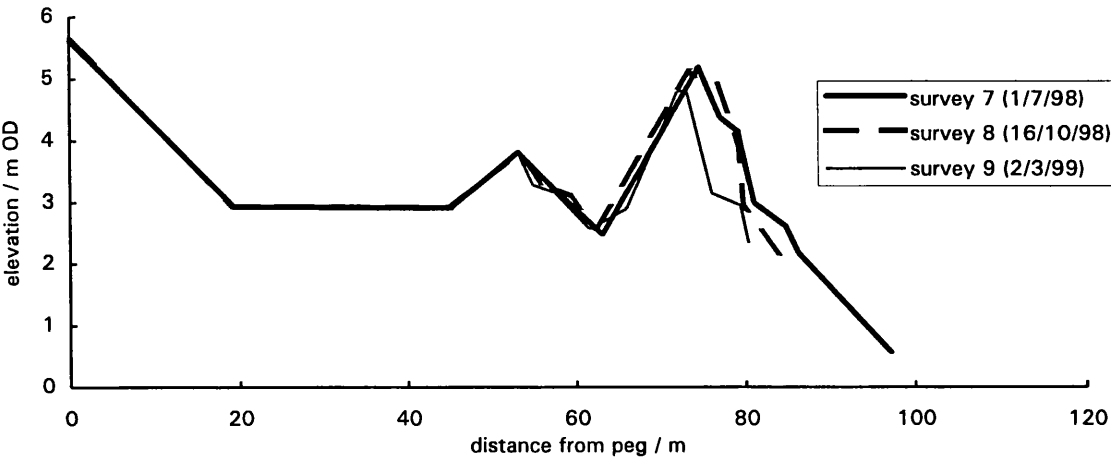
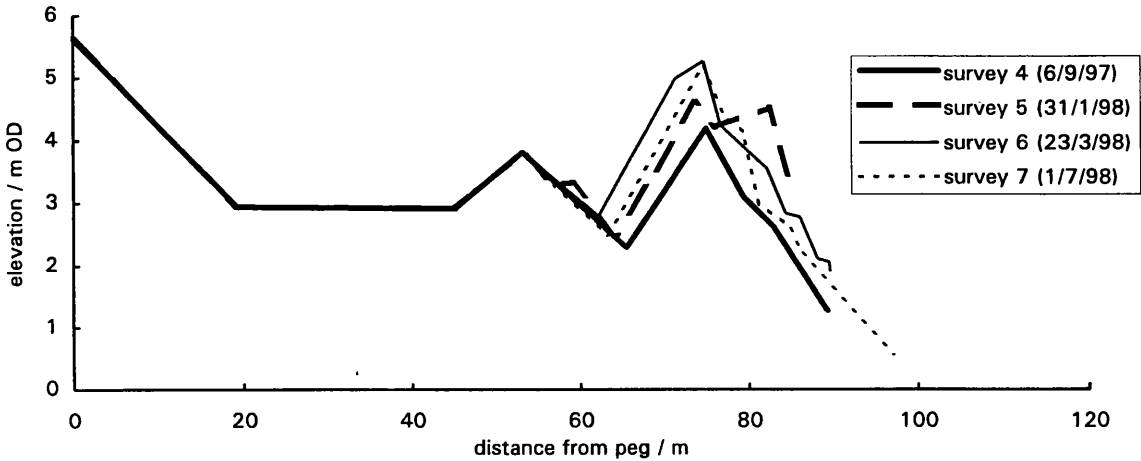
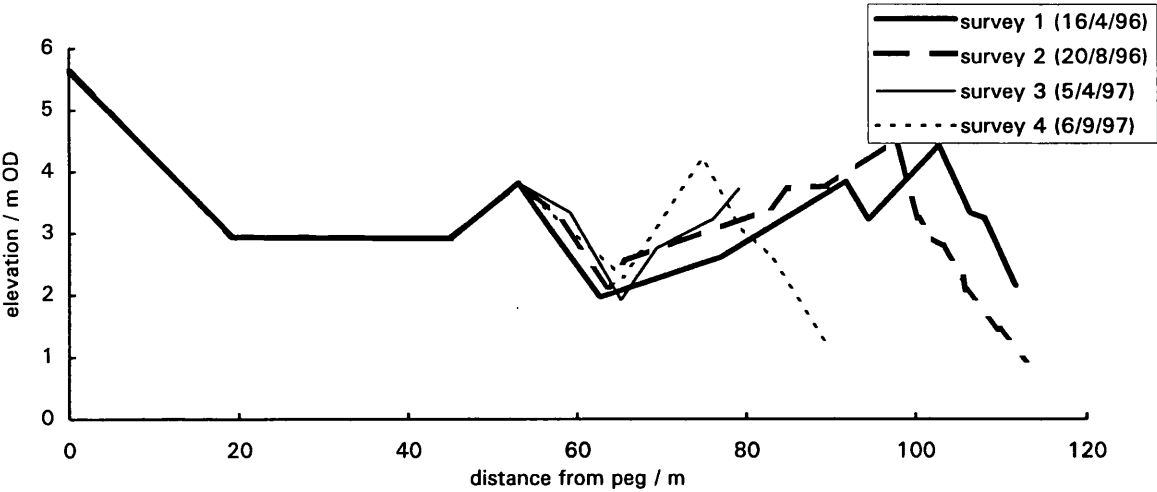
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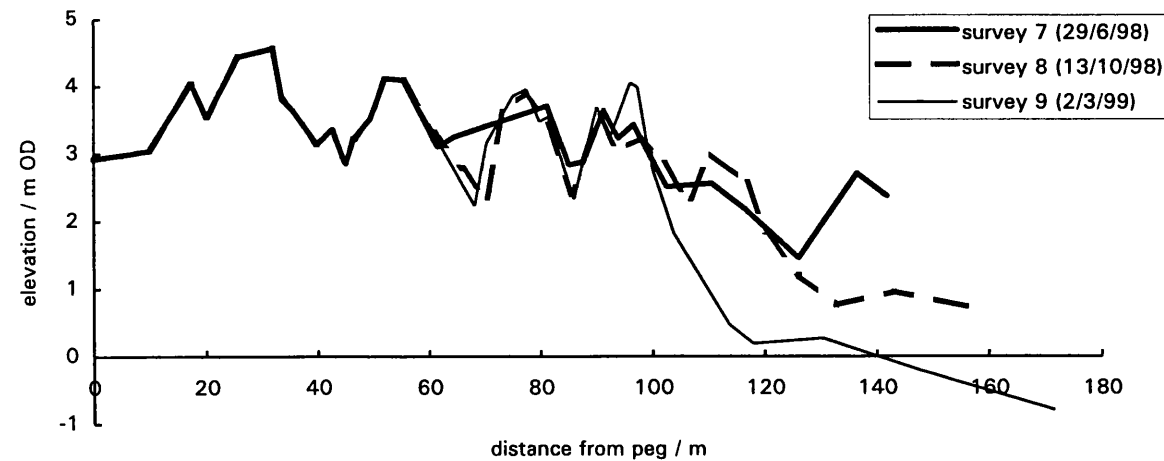
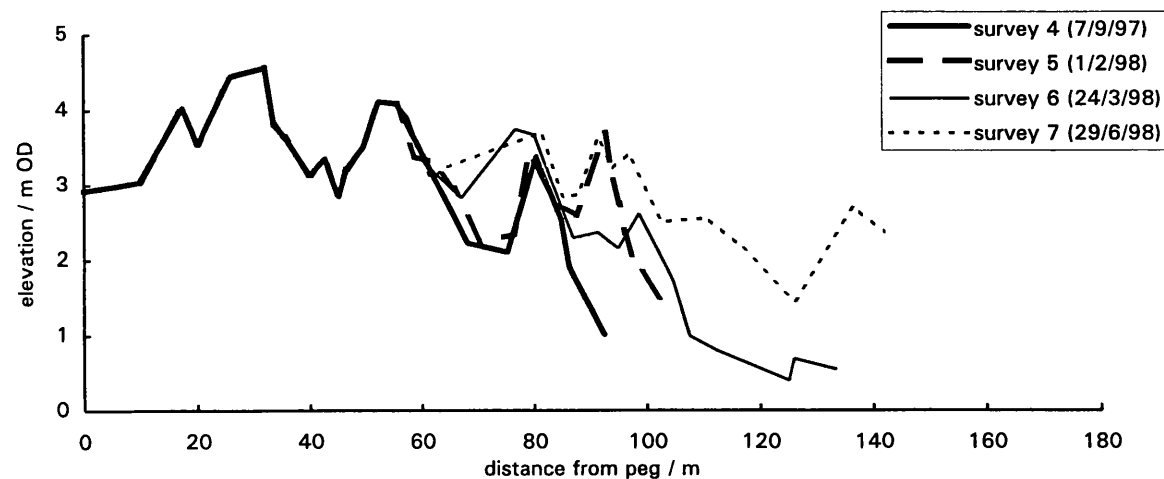
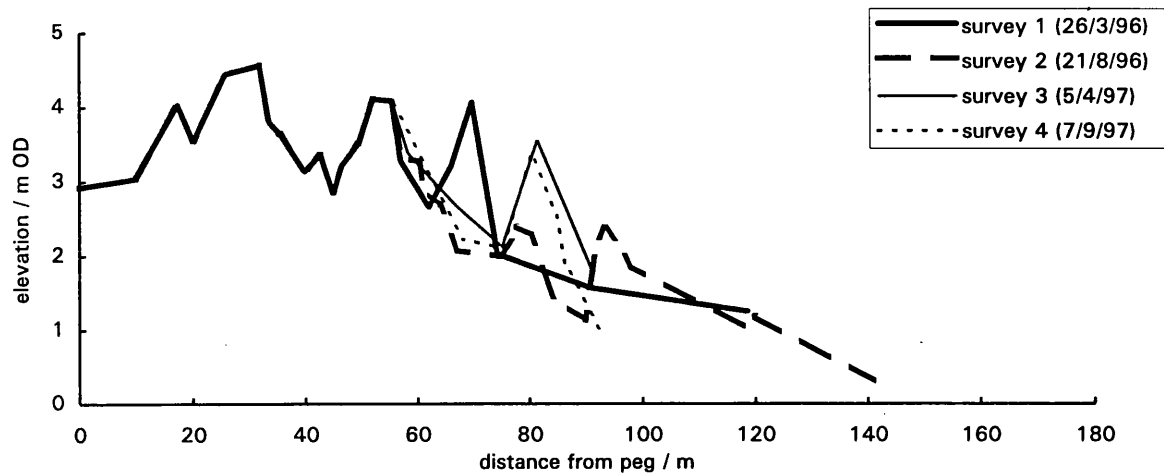
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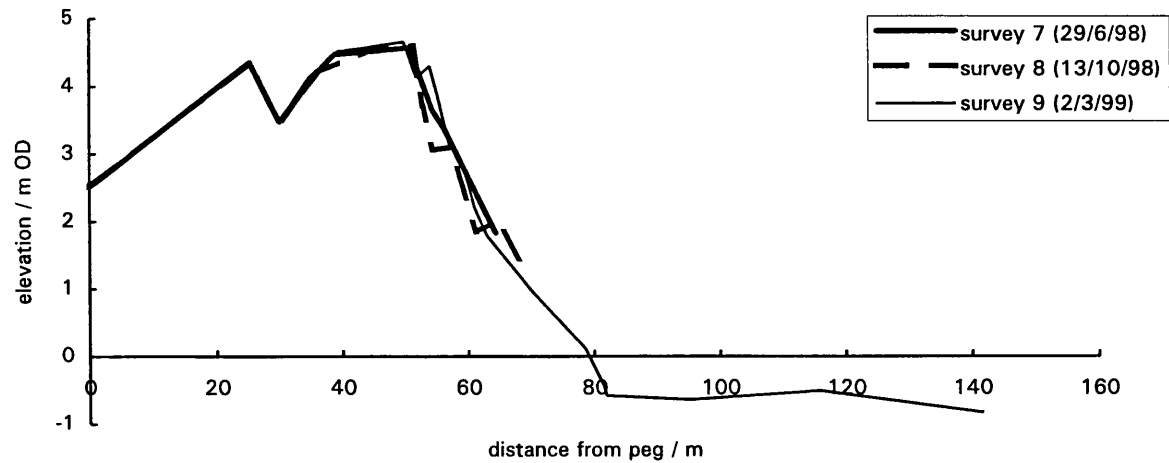
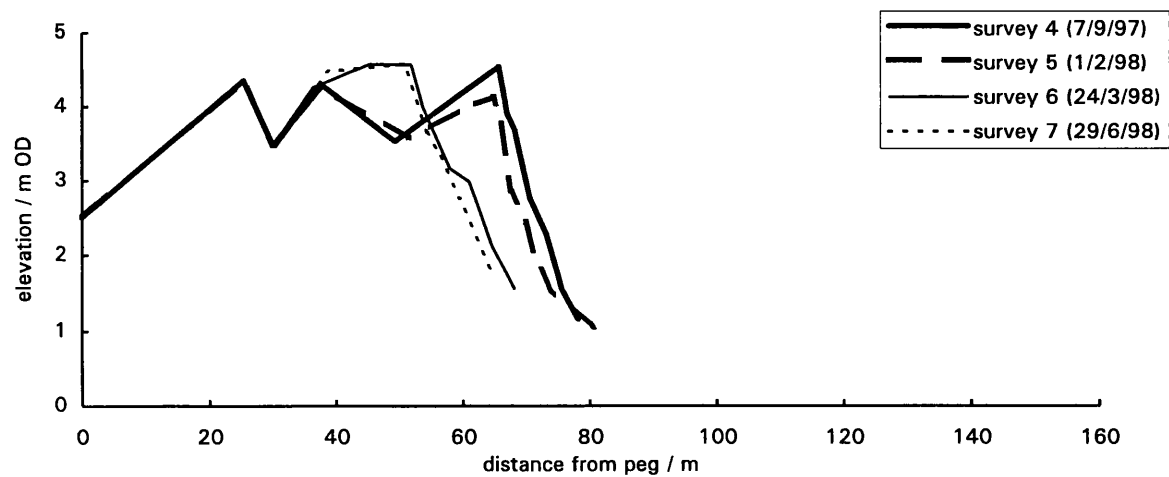
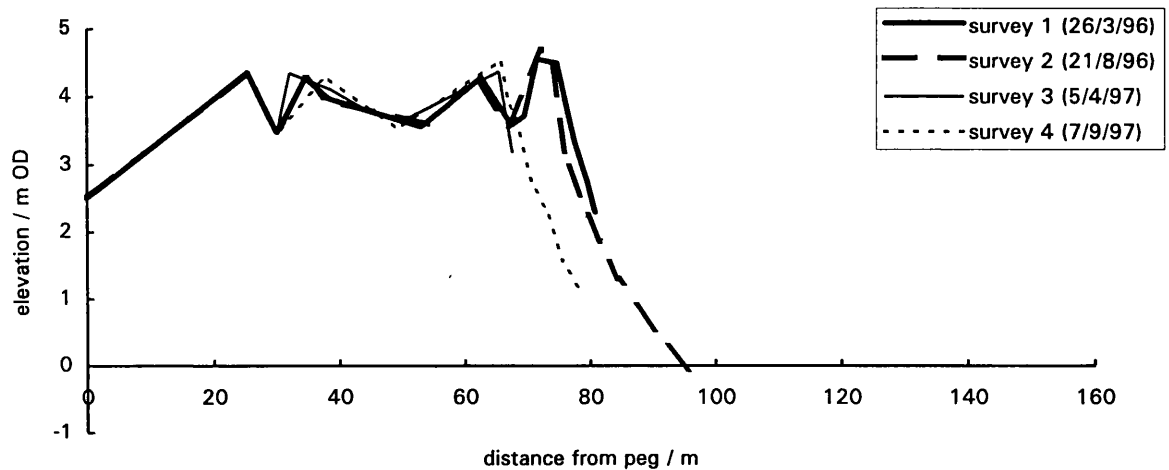
Profile -0.1: Morphological Change (March 1996 to March 1999)



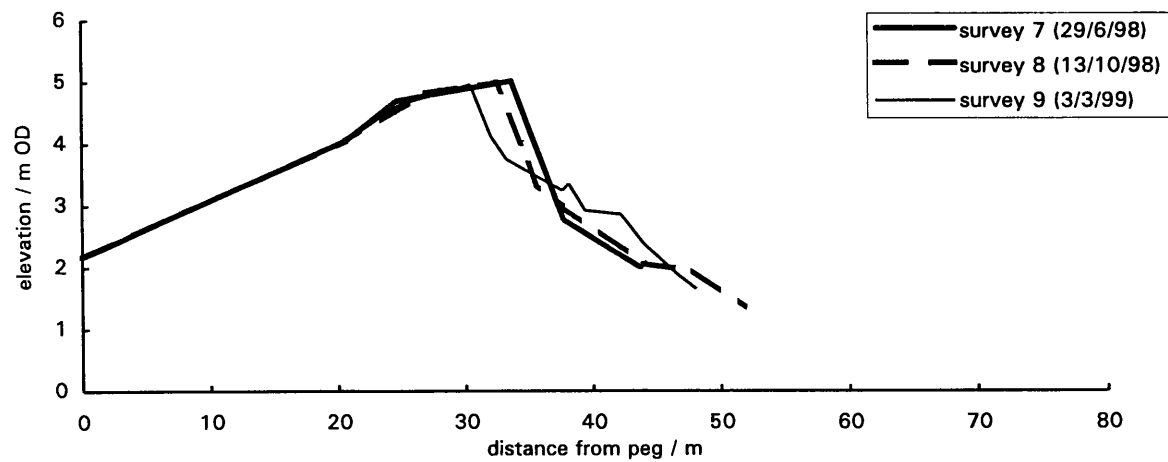
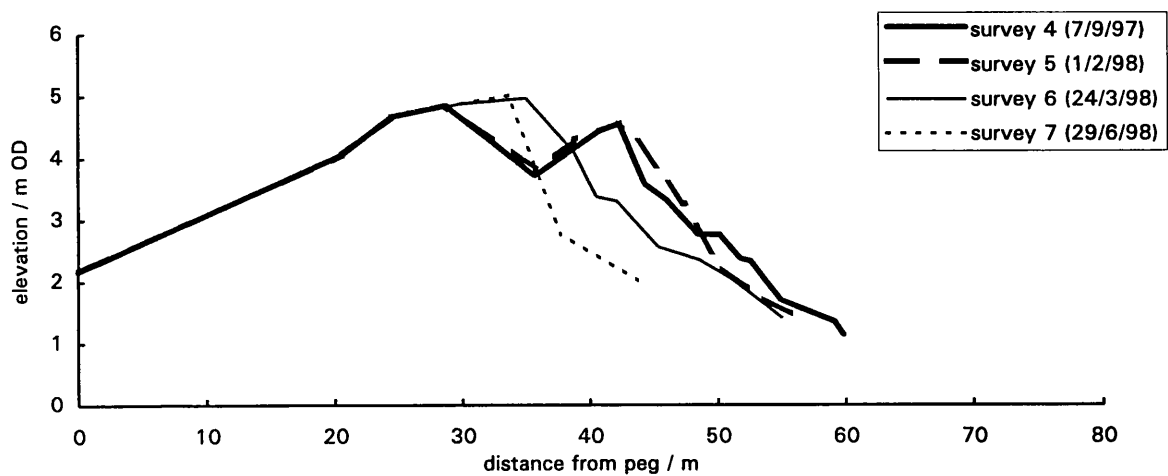
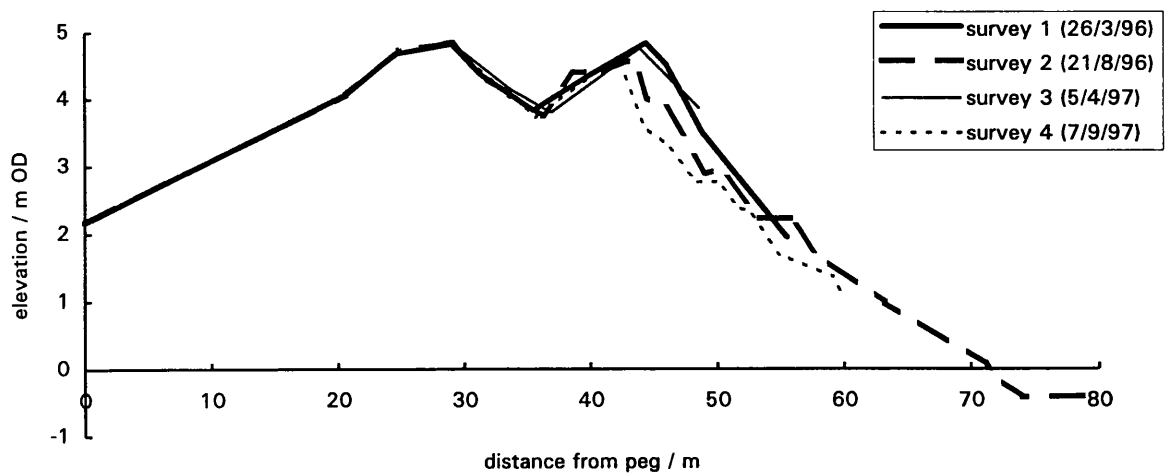
Profile -0.5: Morphological Change (March 1996 to March 1999)



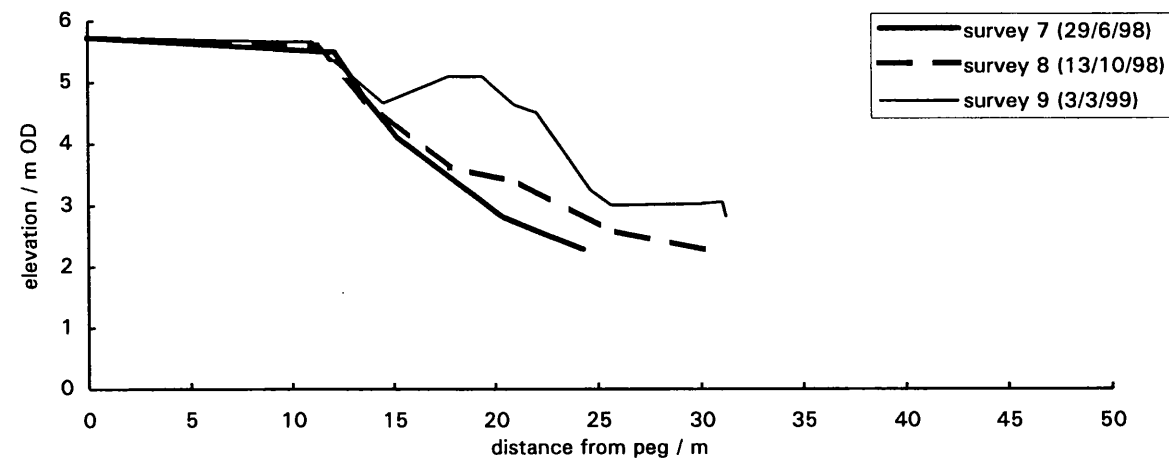
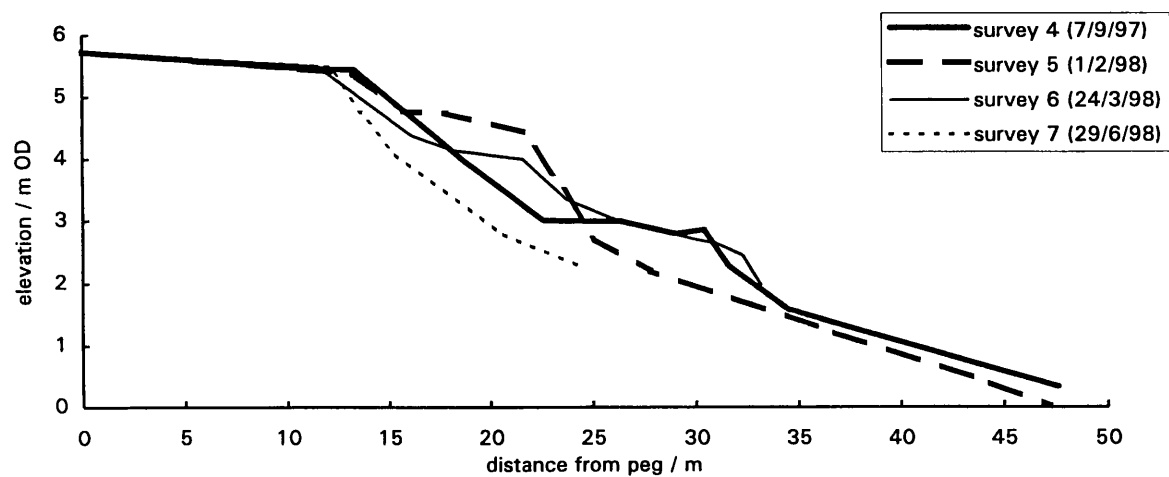
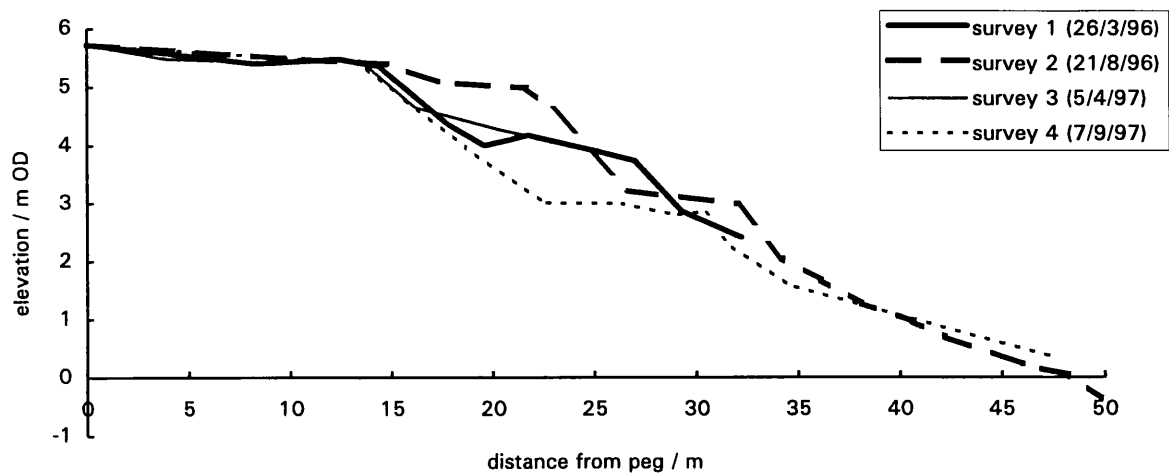
Profile -0.8: Morphological Change (March 1996 to March 1999)



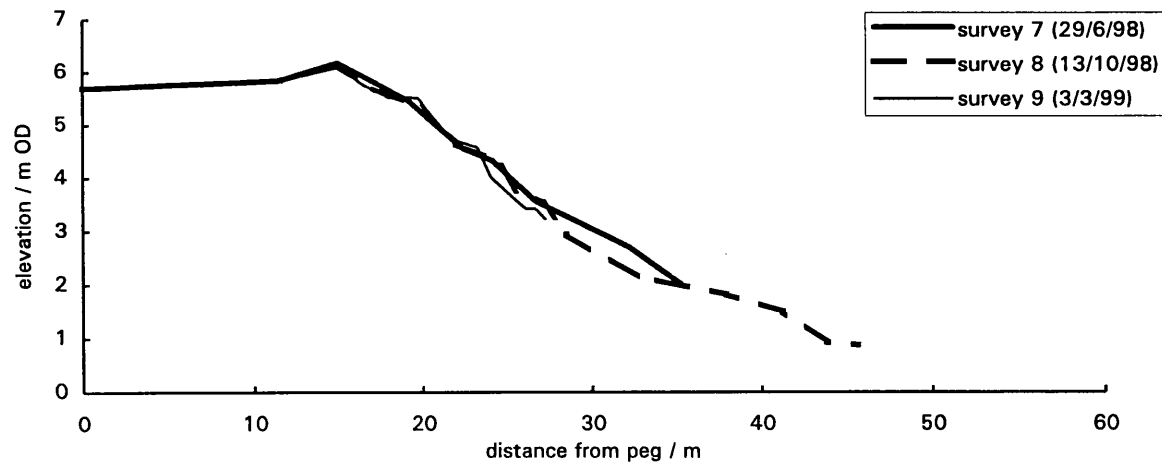
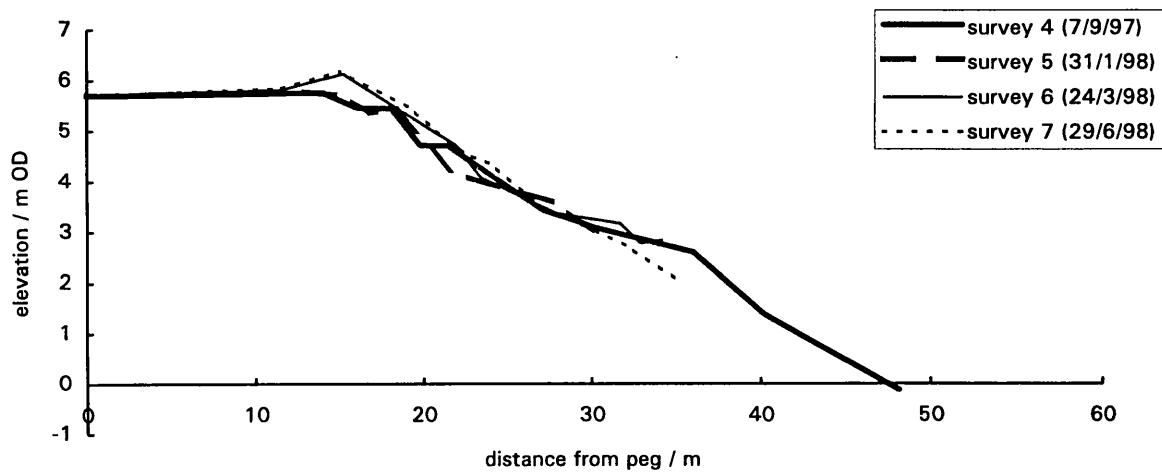
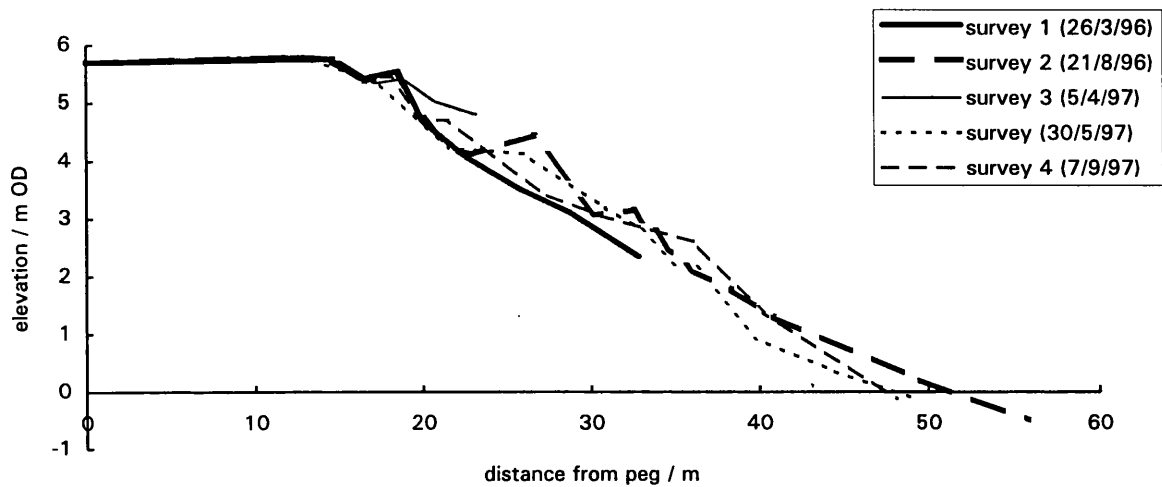
Profile -1: Morphological Change (April 1996 to March 1999)



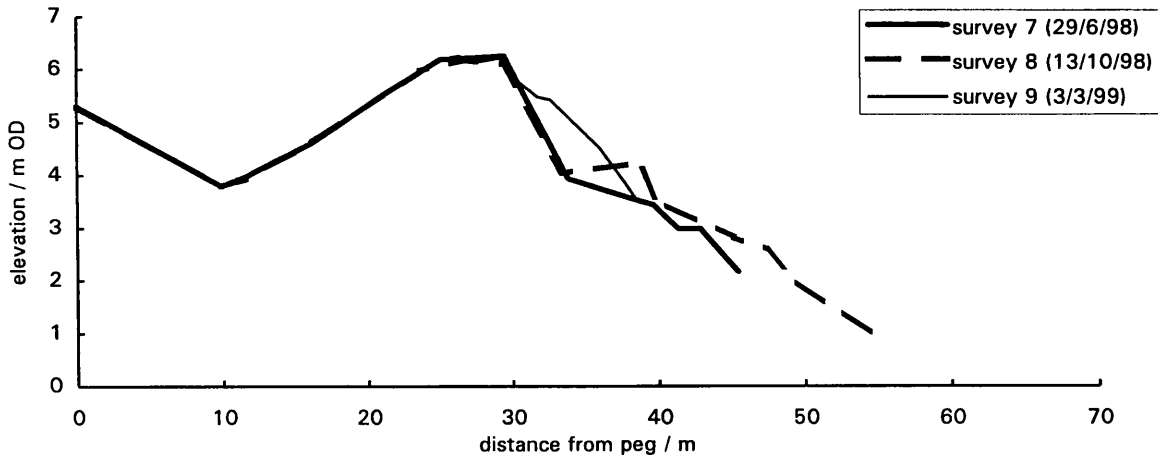
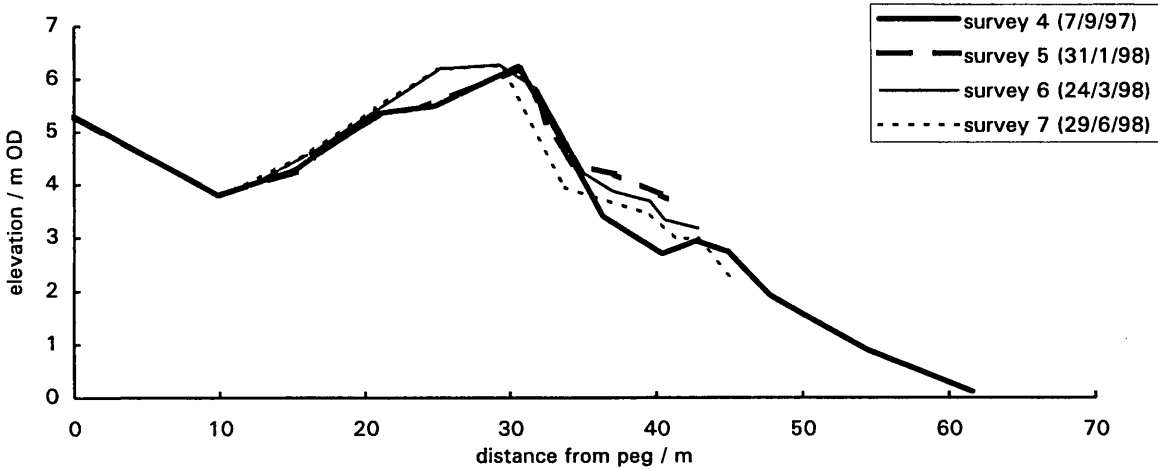
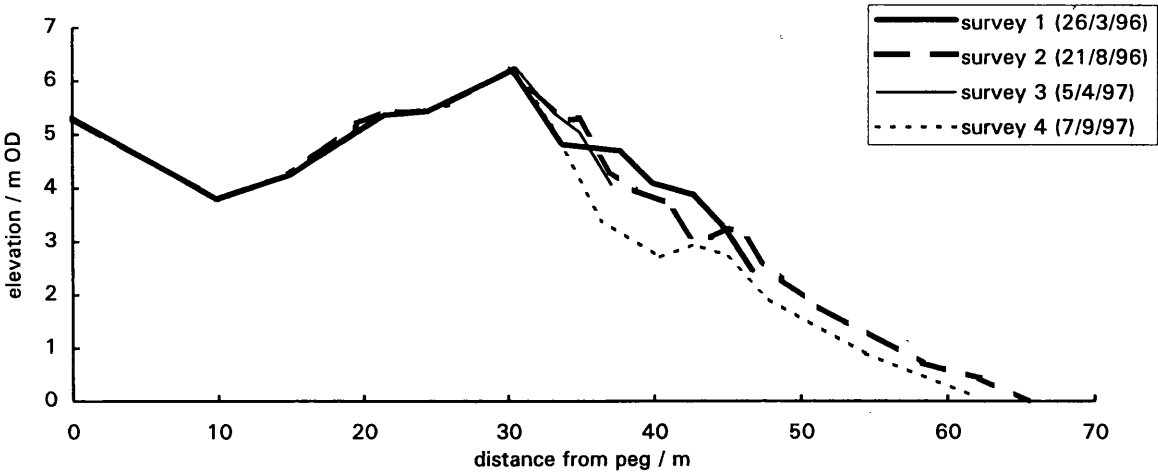
Profile -1.5: Morphological Change (March 1996 to March 1999)



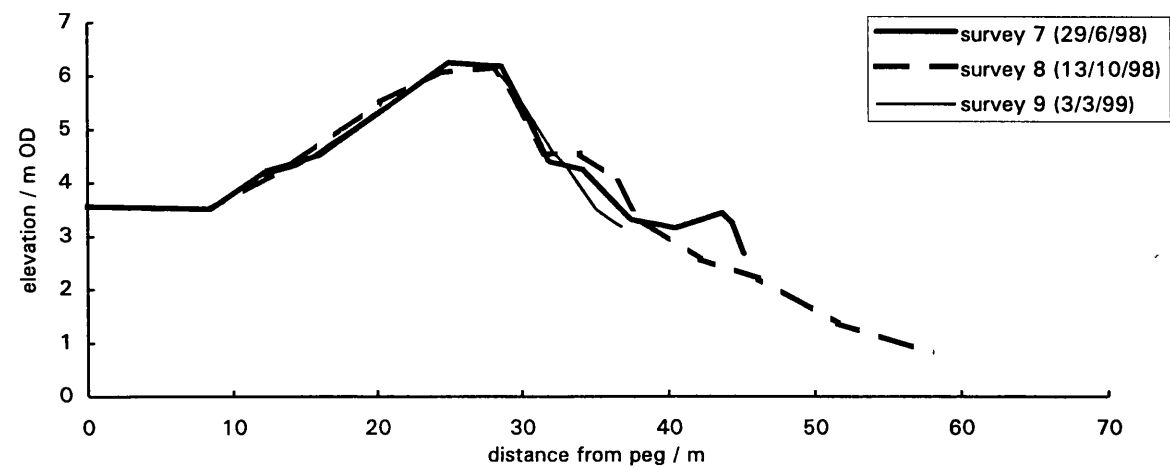
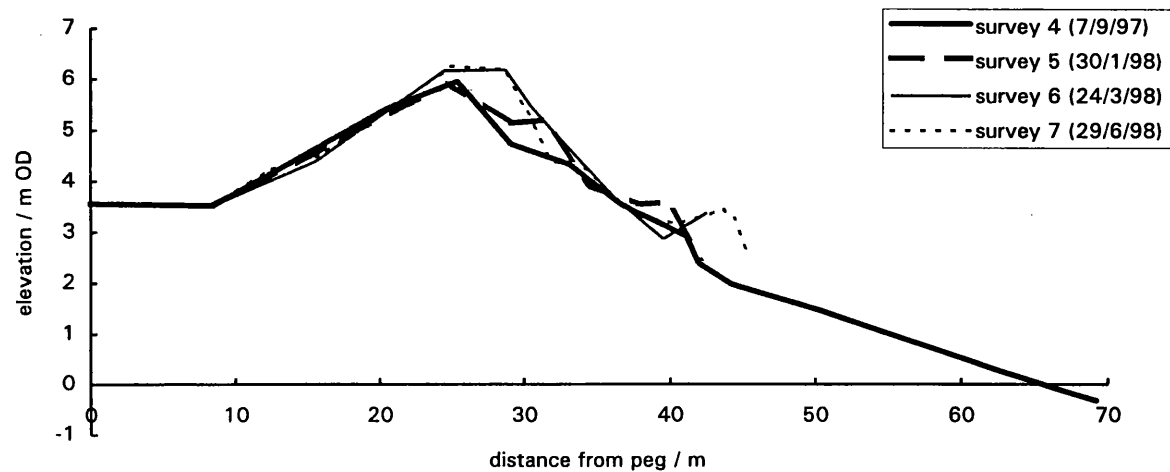
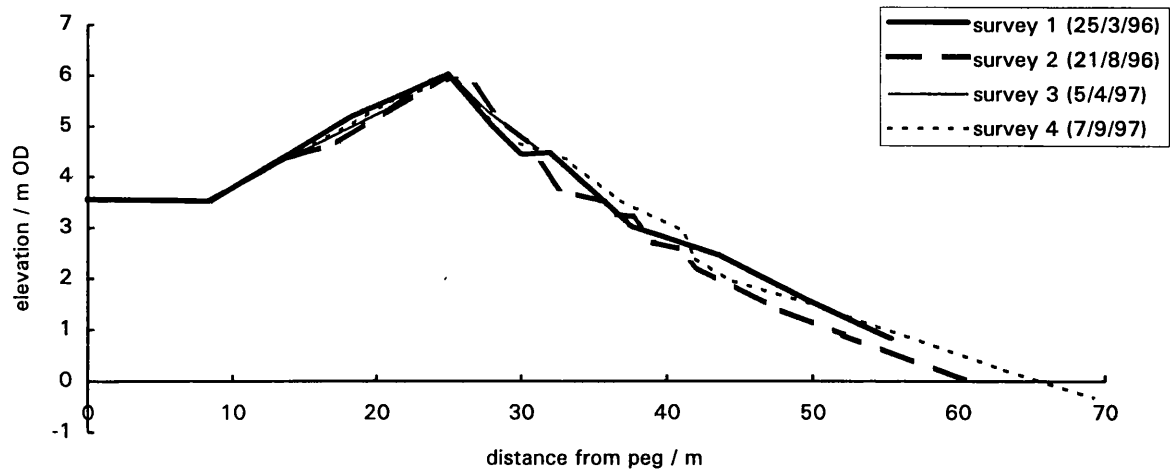
Profile -2: Morphological Change (March 1996 to March 1999)



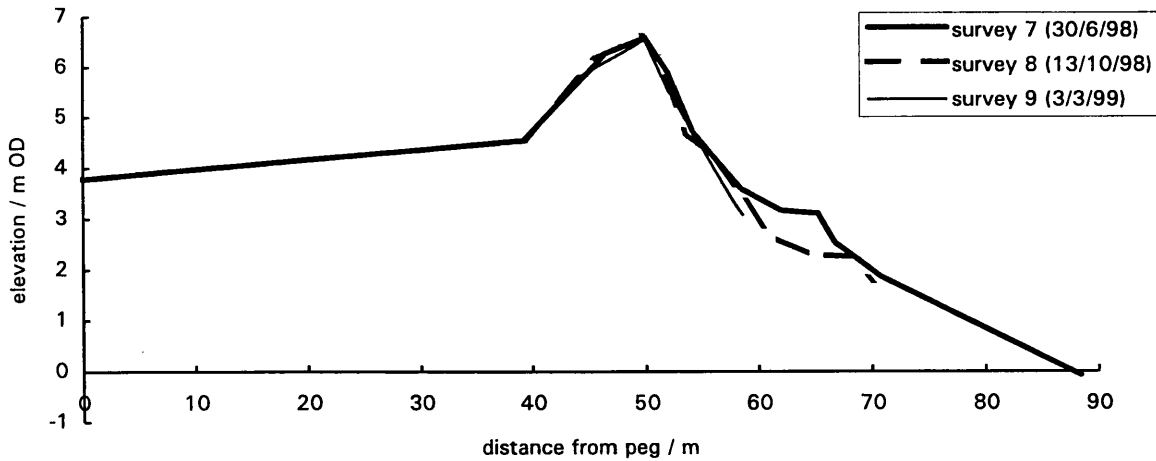
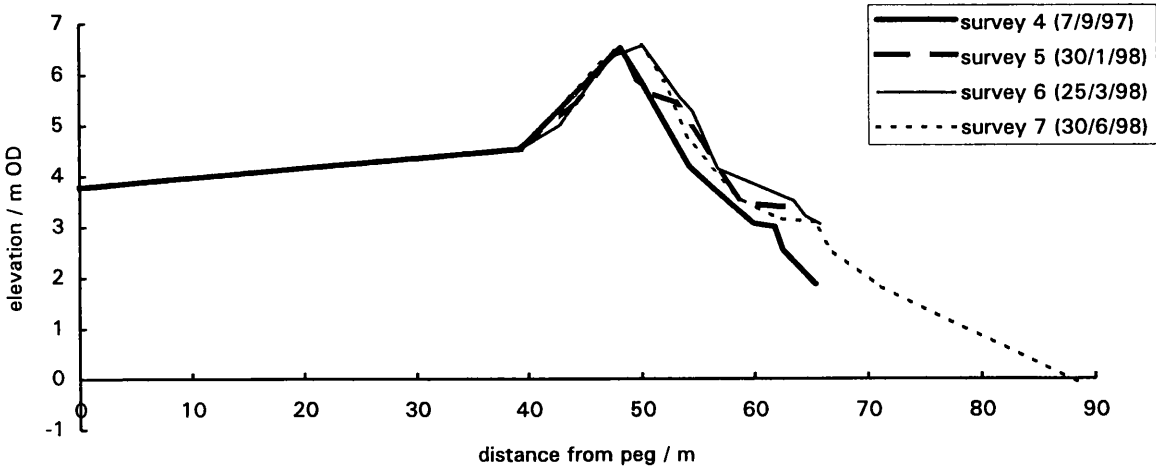
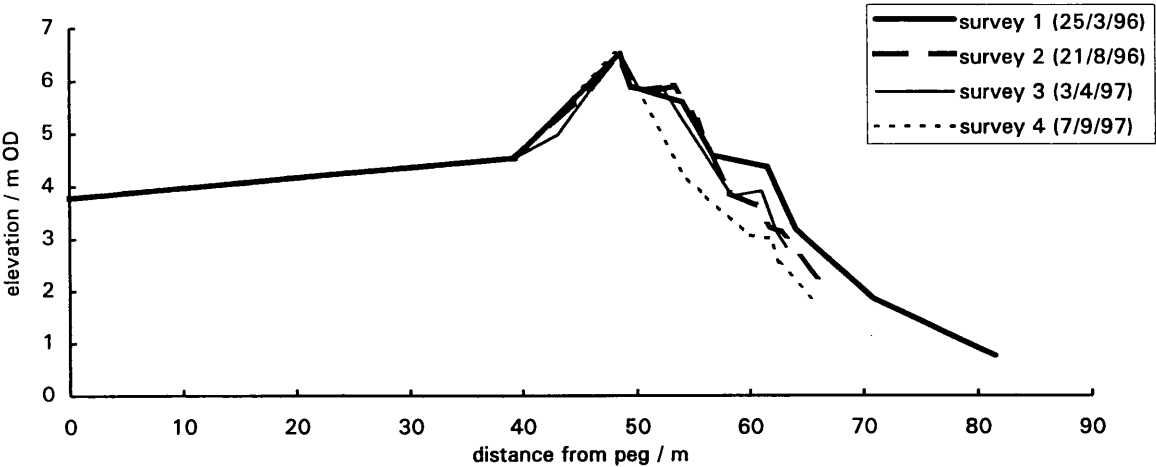
Profile -2.5: Morphological Change (March 1996 to March 1999)



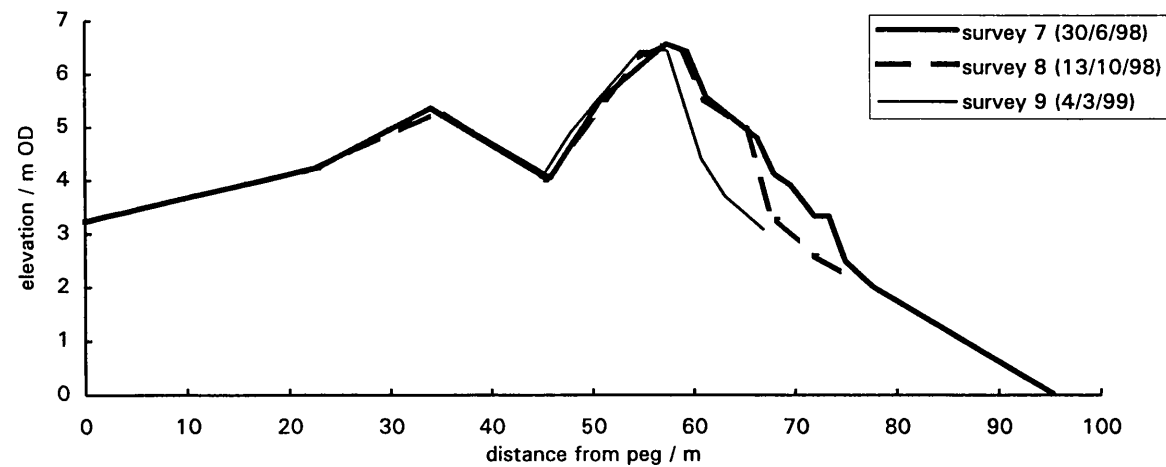
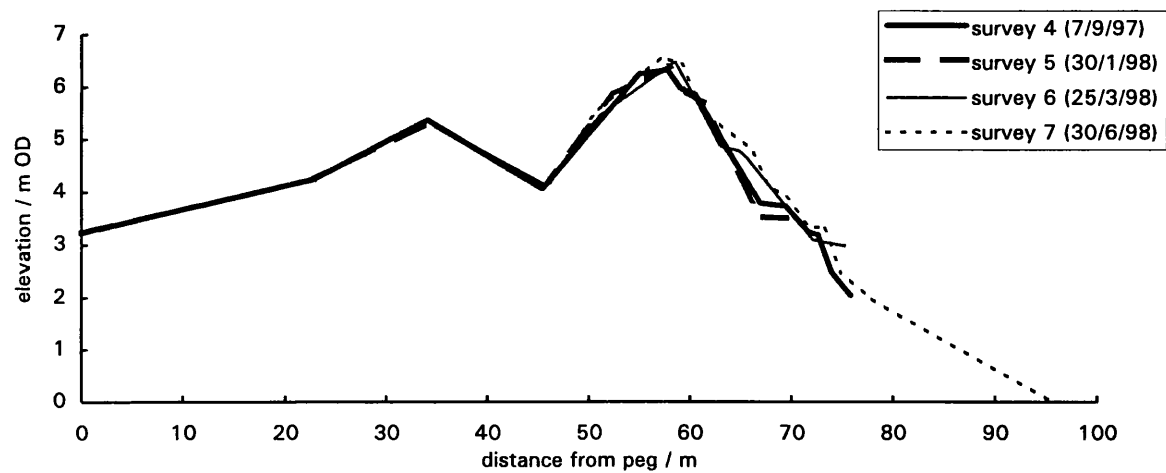
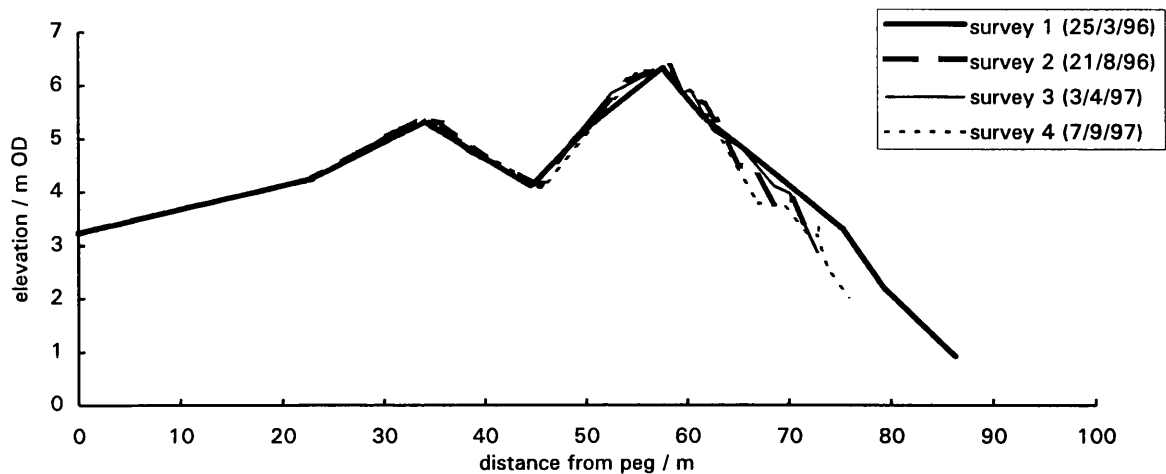
Profile -3: Morphological Change (March 1996 to March 1999)



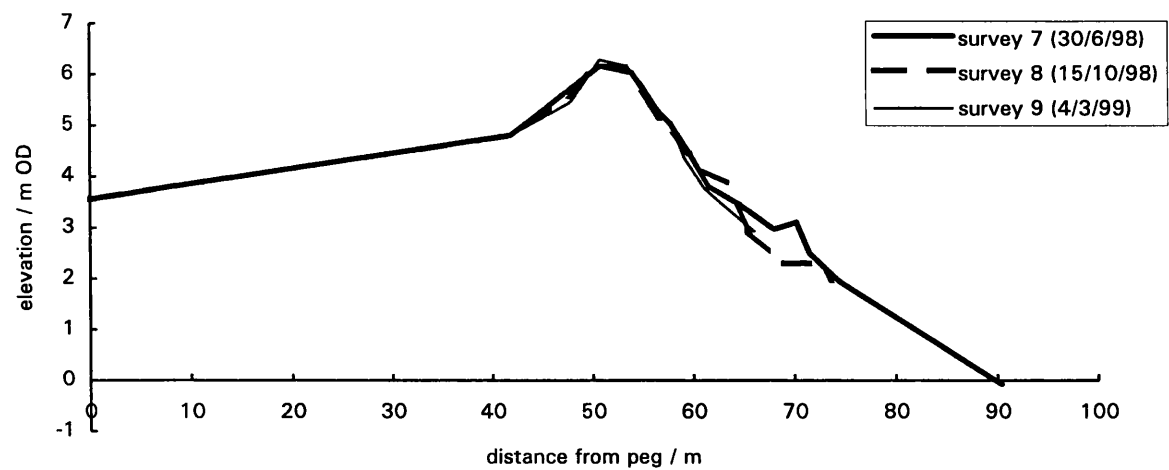
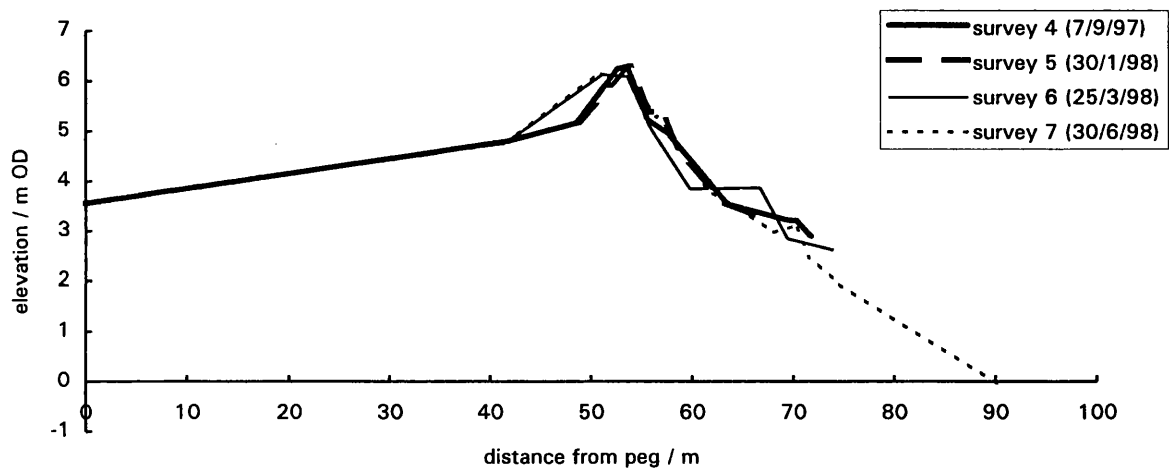
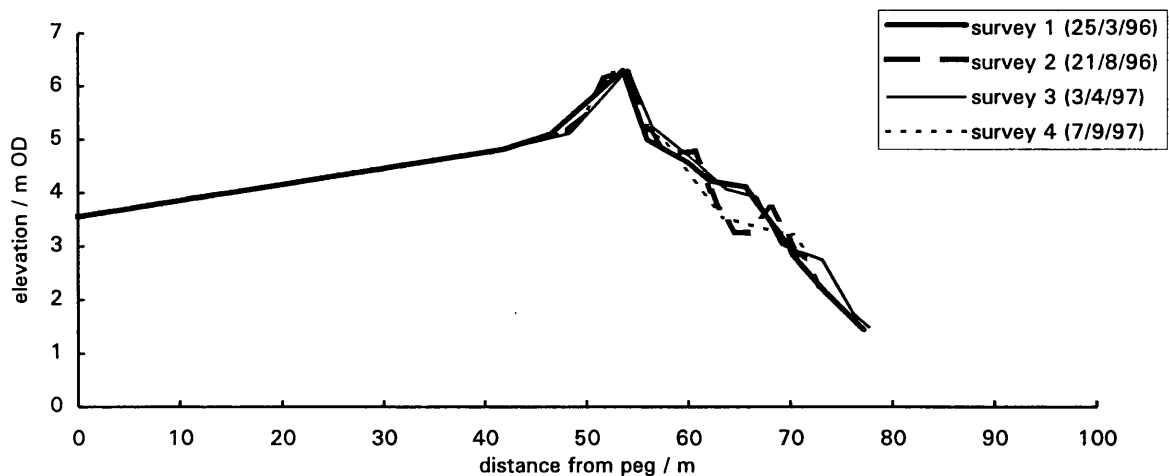
Profile -3.5: Morphological Change (March 1996 to March 1999)



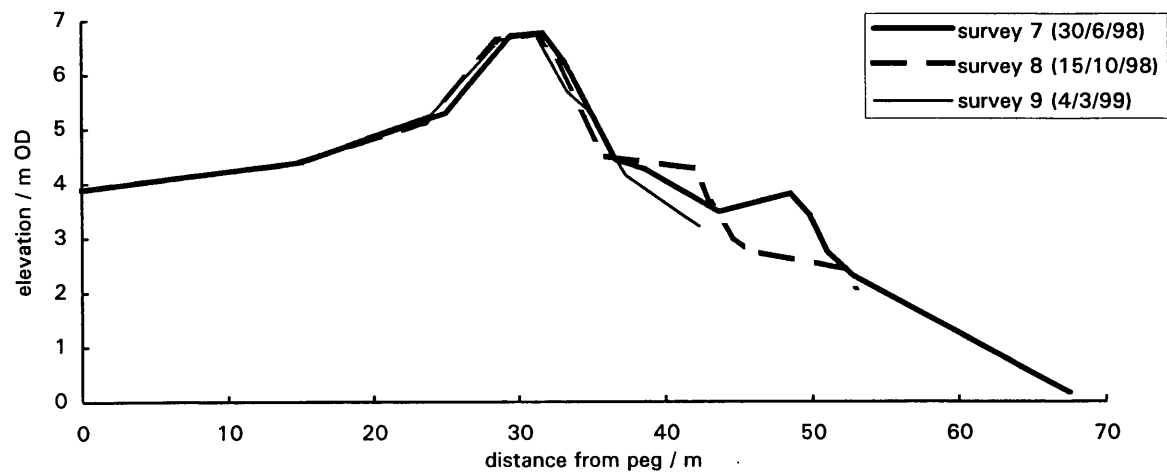
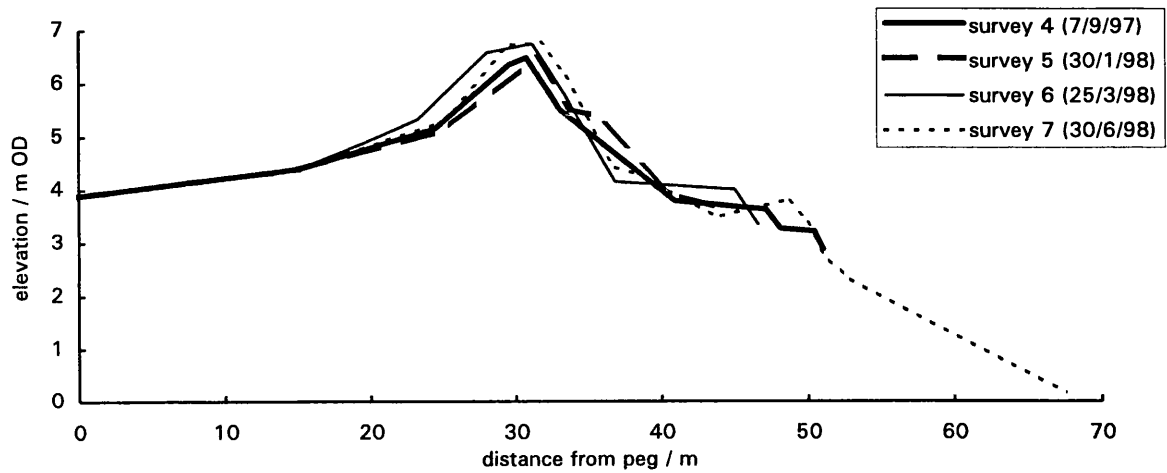
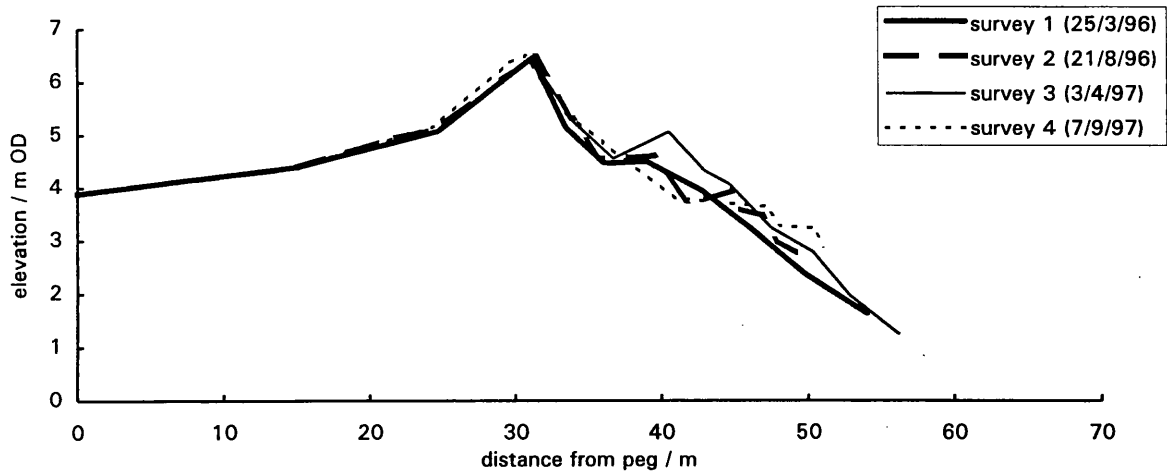
Profile -4: Morphological Change (March 1996 to March 1999)



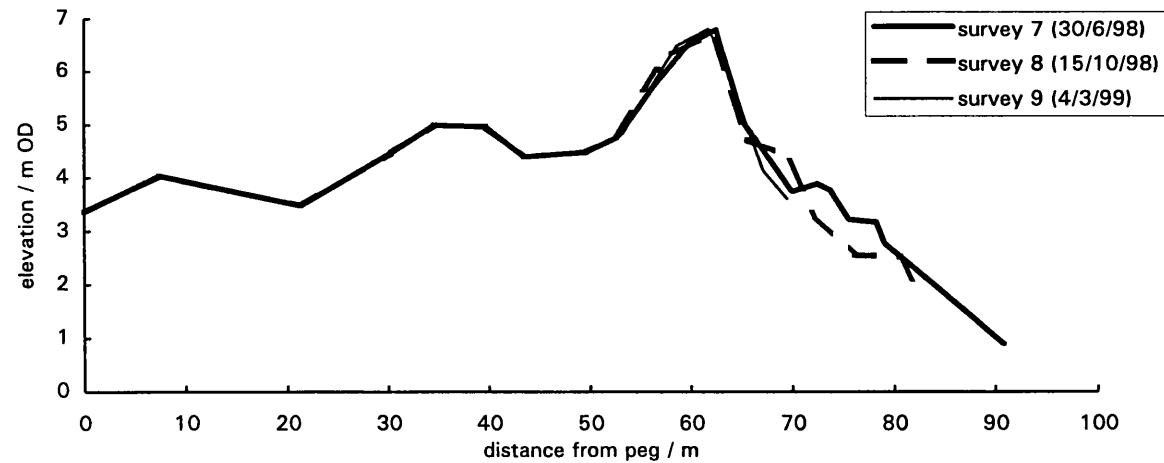
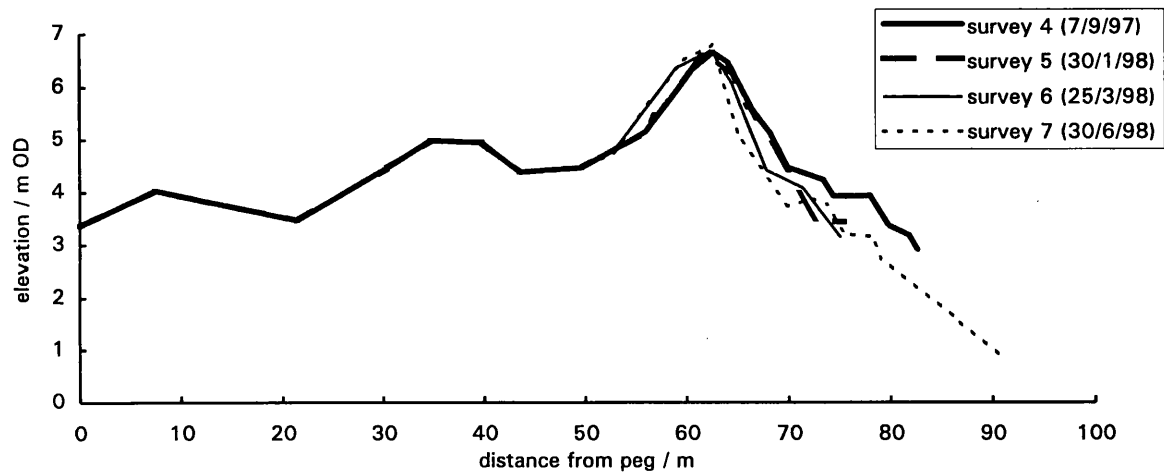
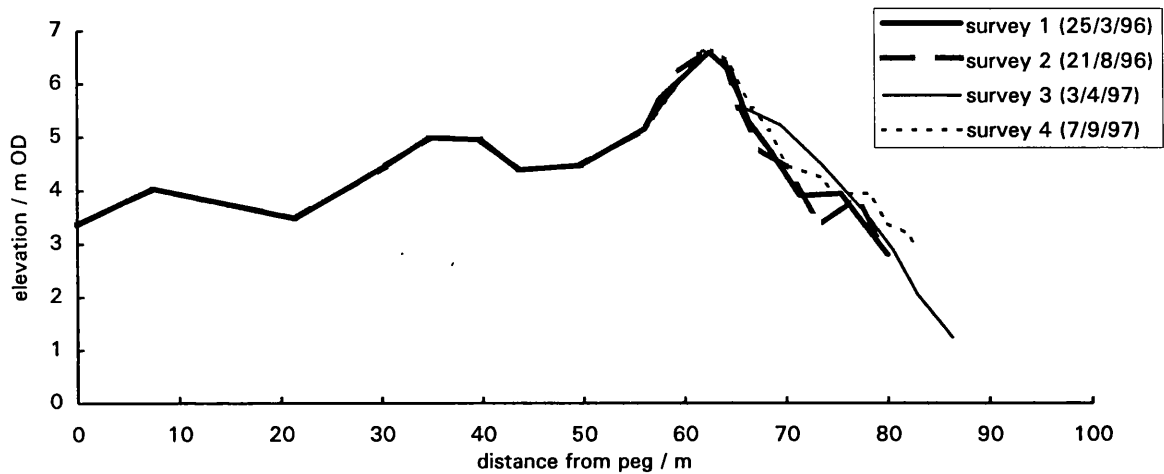
Profile -4.5: Morphological Change (March 1996 to March 1999)



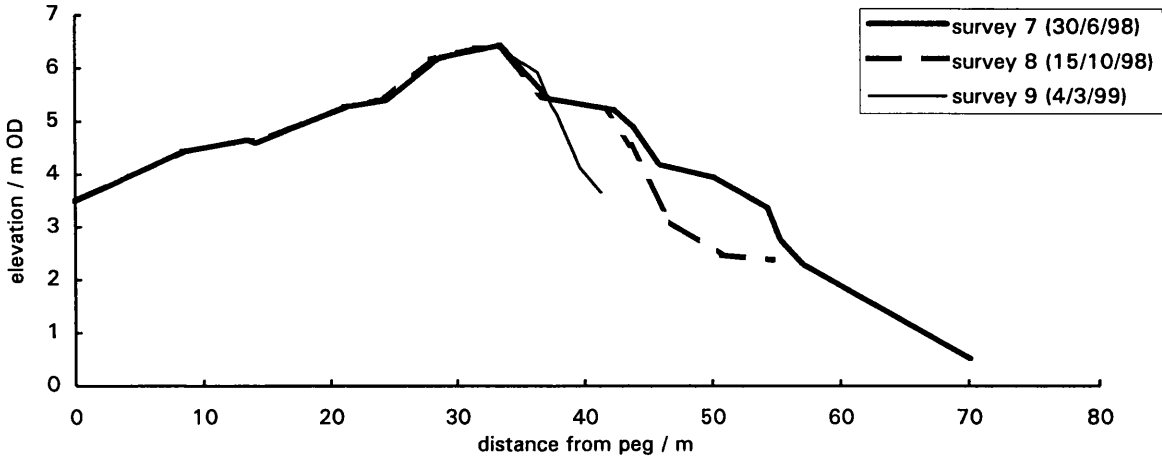
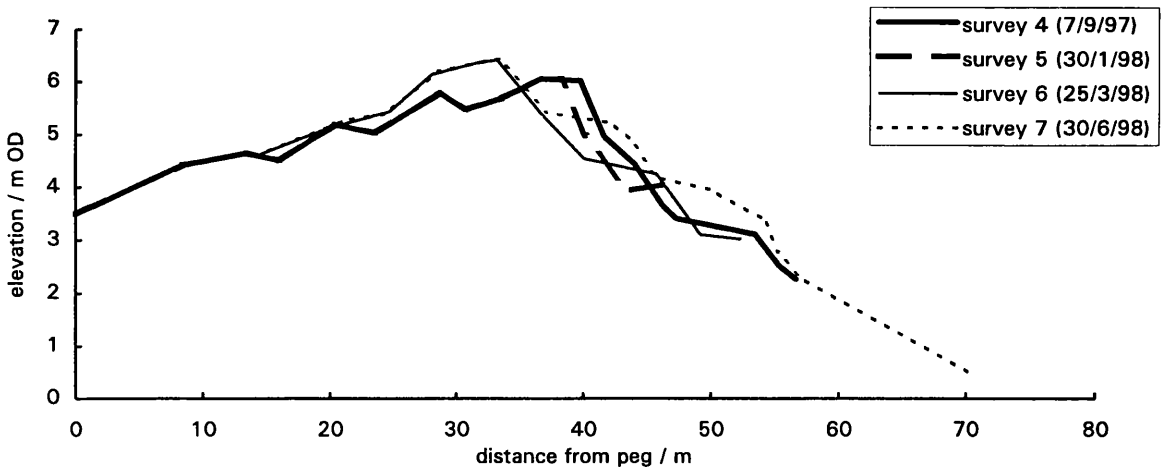
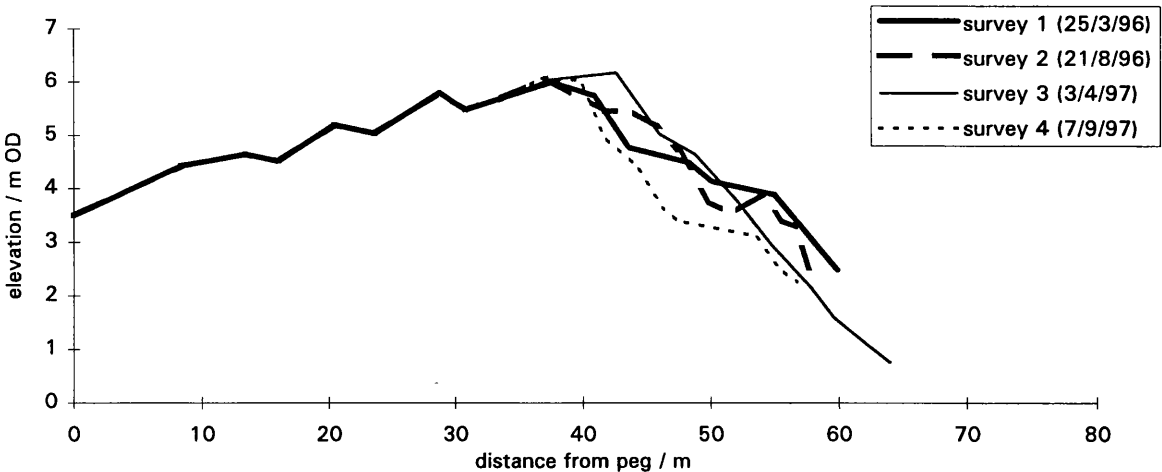
Profile -5: Morphological Change (March 1996 to March 1999)



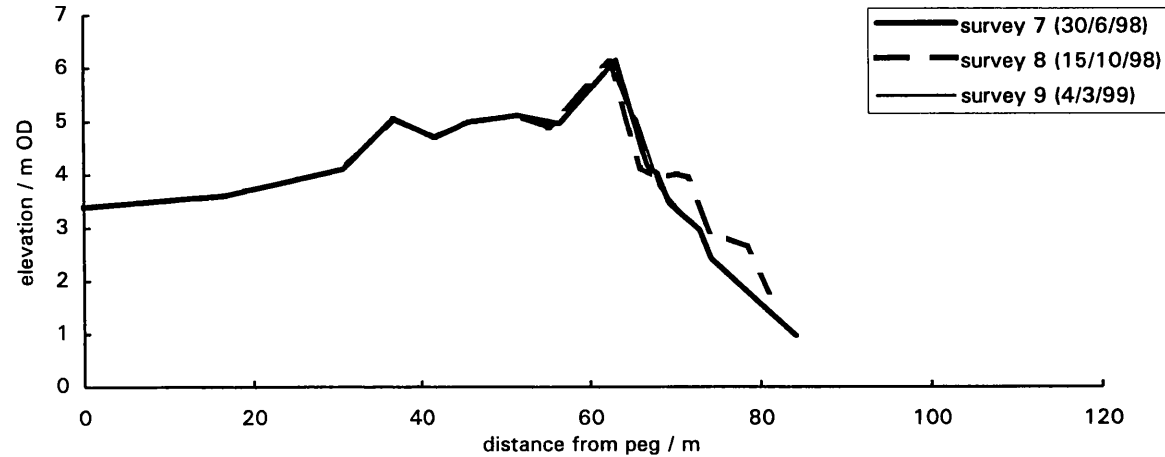
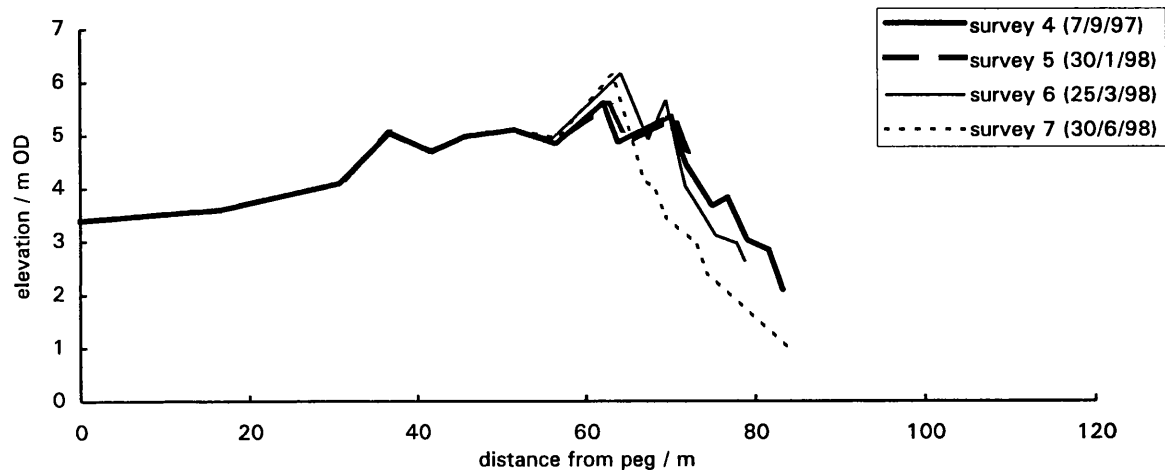
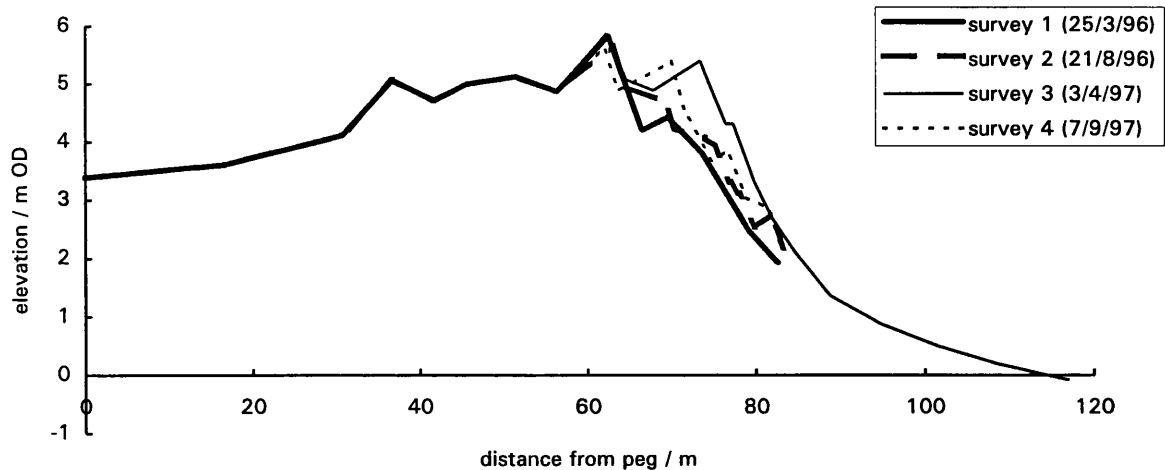
Profile -5.5: Morphological Change (March 1996 to March 1999)



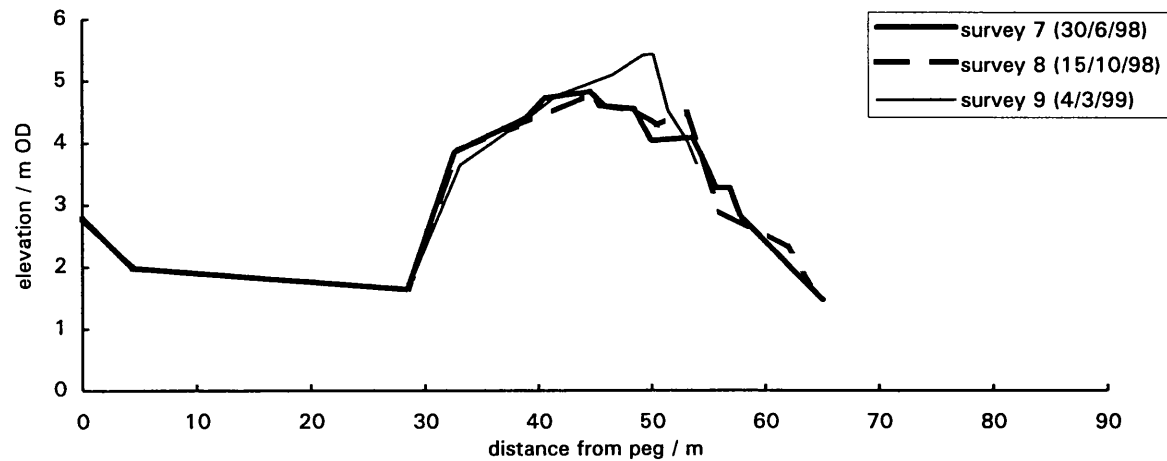
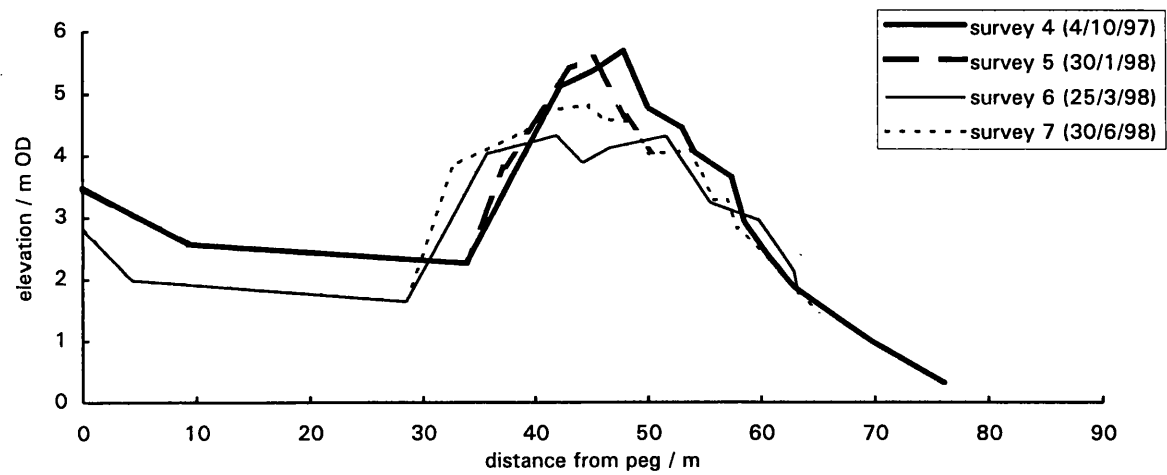
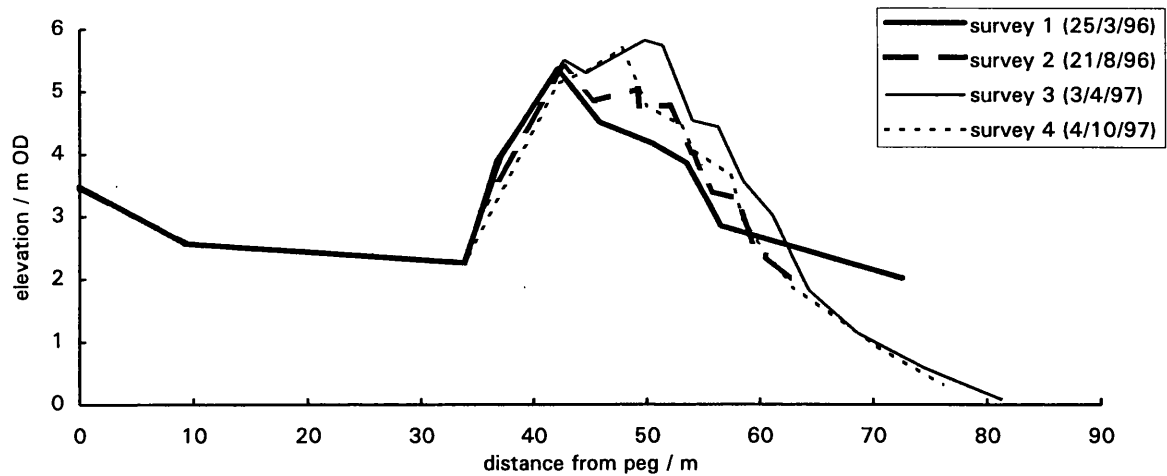
Profile -6: Morphological Change (March 1996 to March 1999)



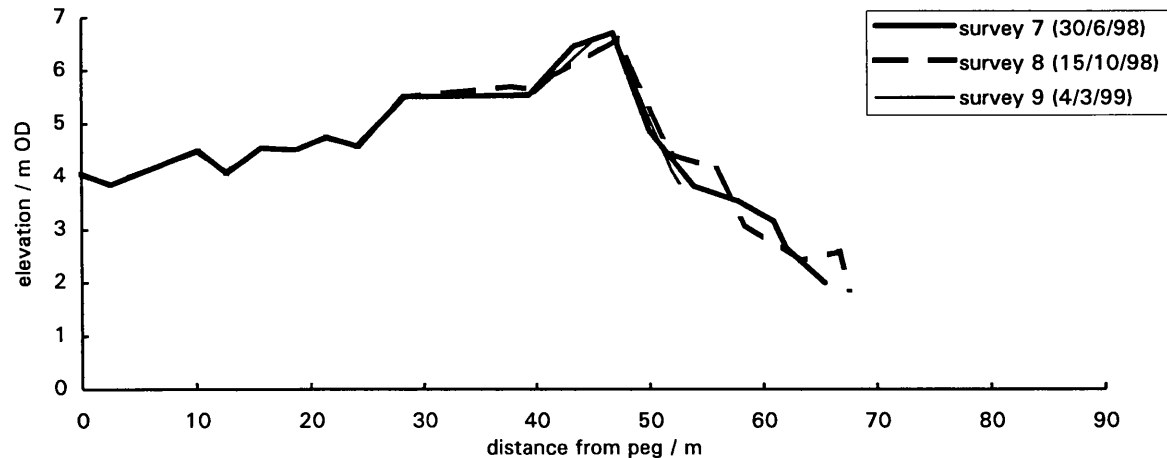
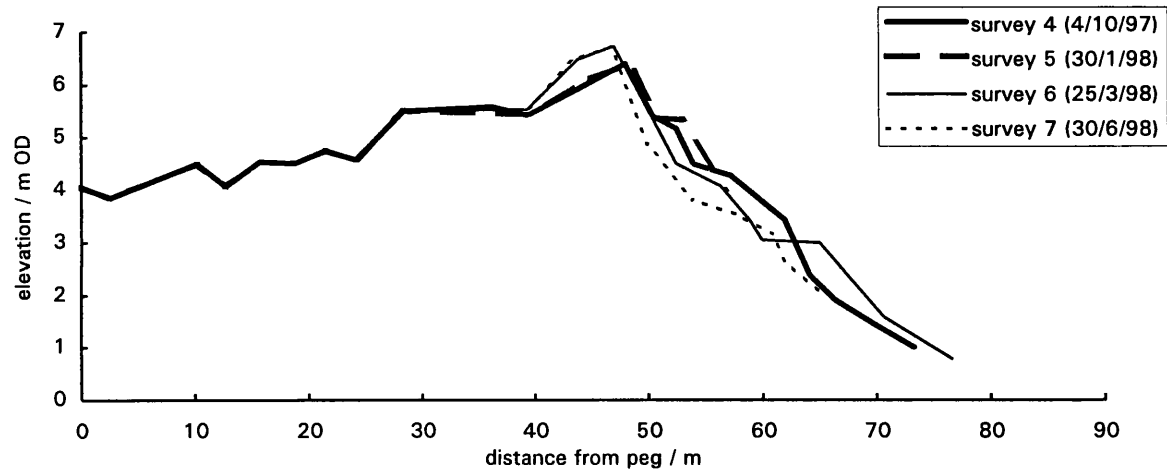
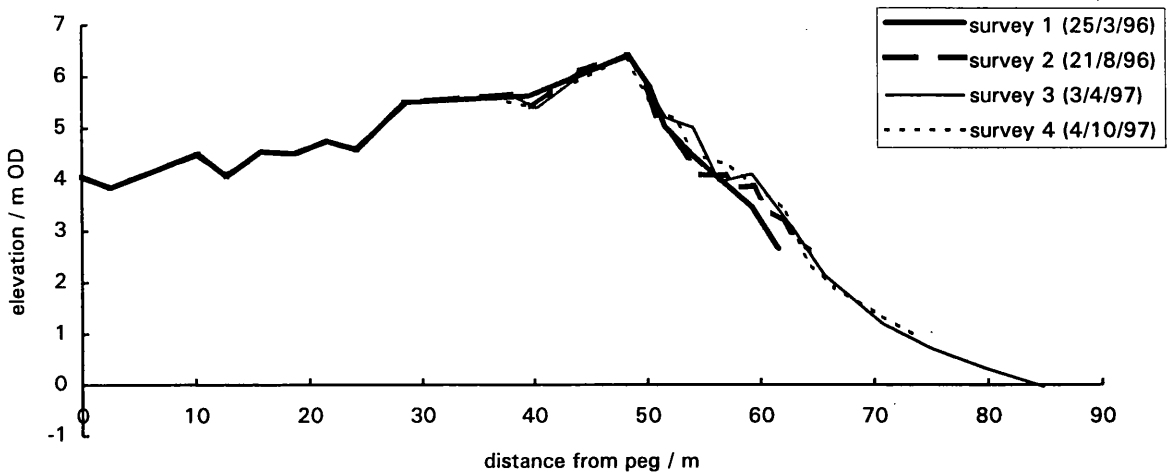
Profile -6.5: Morphological Change (March 1996 to March 1999)



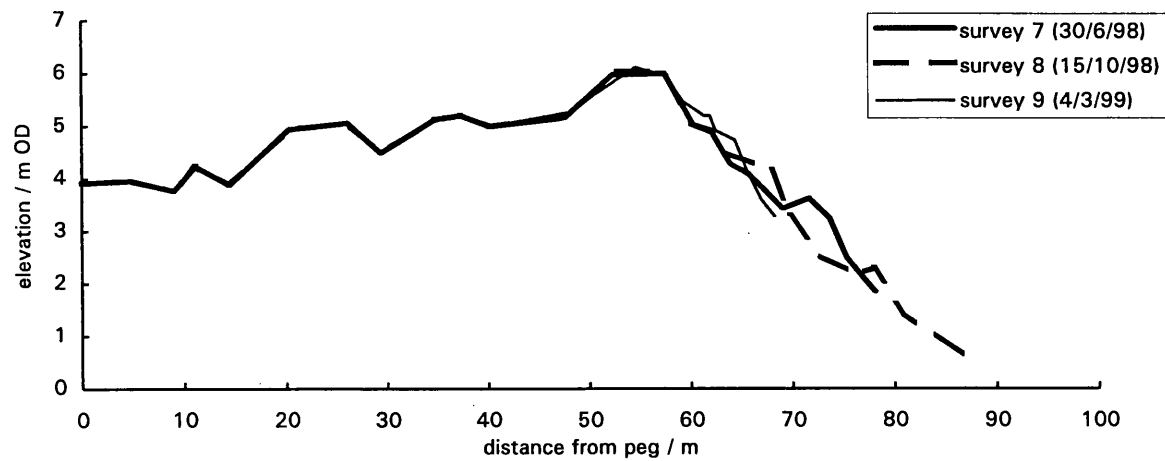
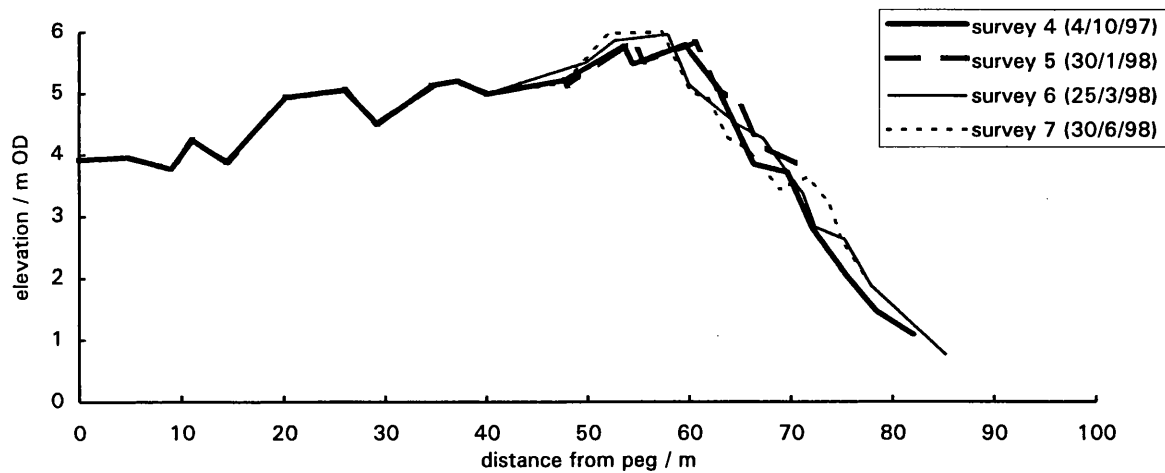
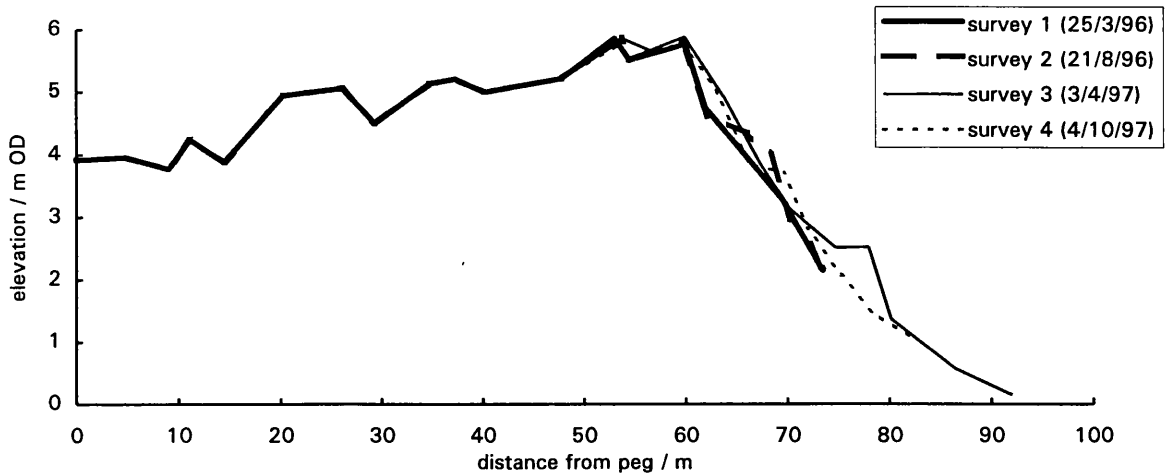
Profile -7: Morphological Change (March 1996 to March 1999)



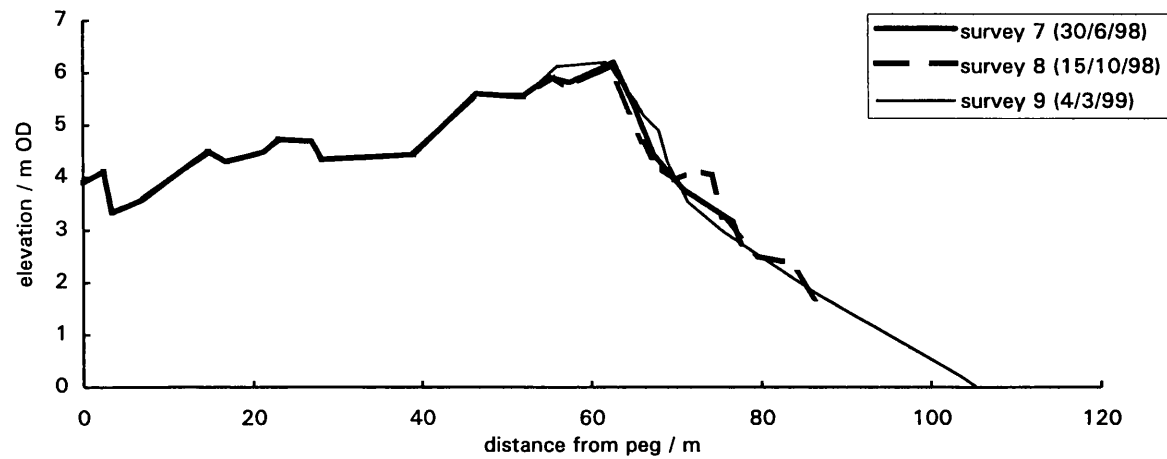
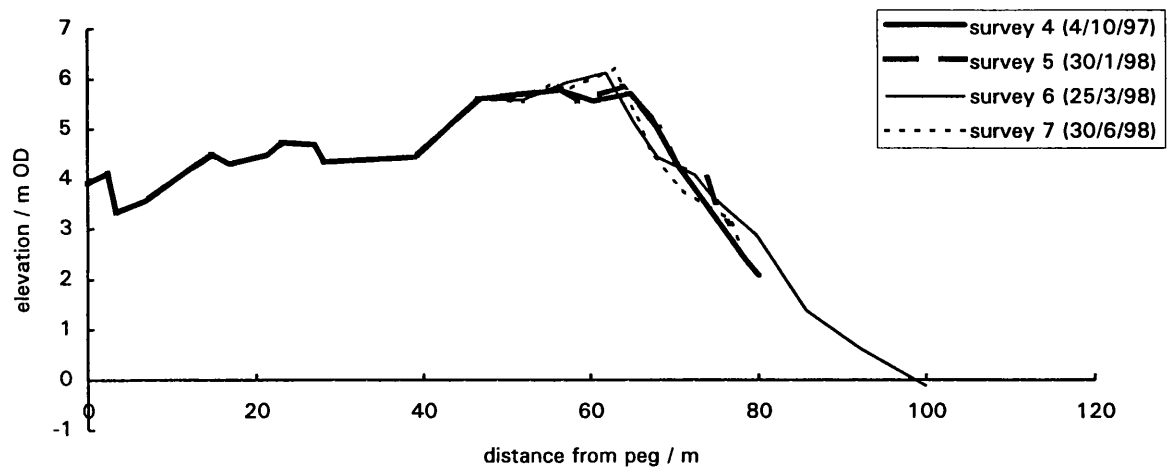
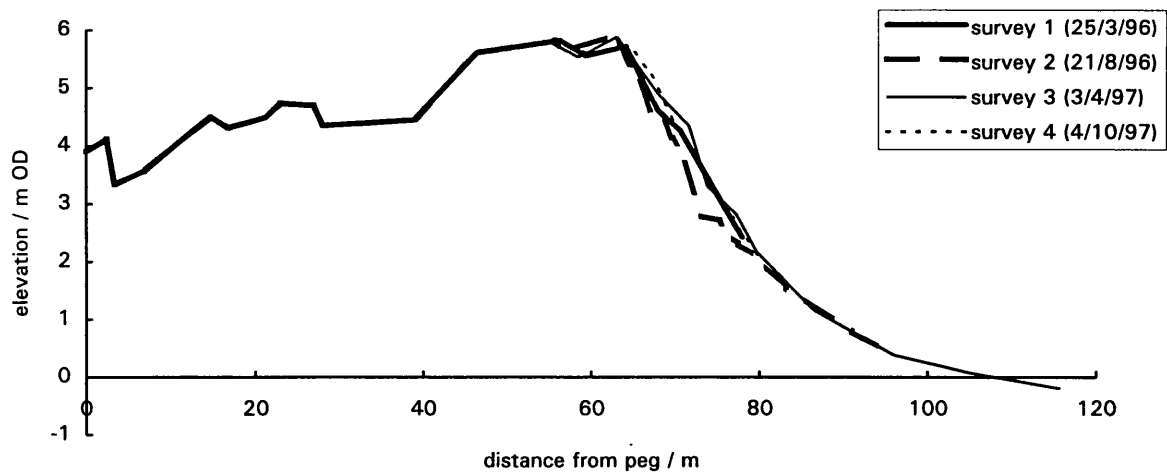
Profile -7.5: Morphological Change (March 1996 to March 1999)



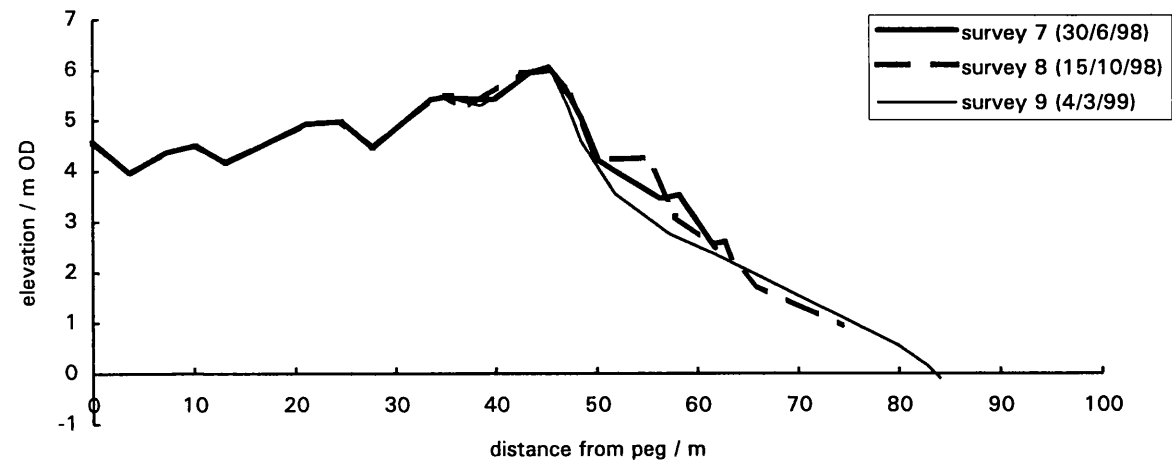
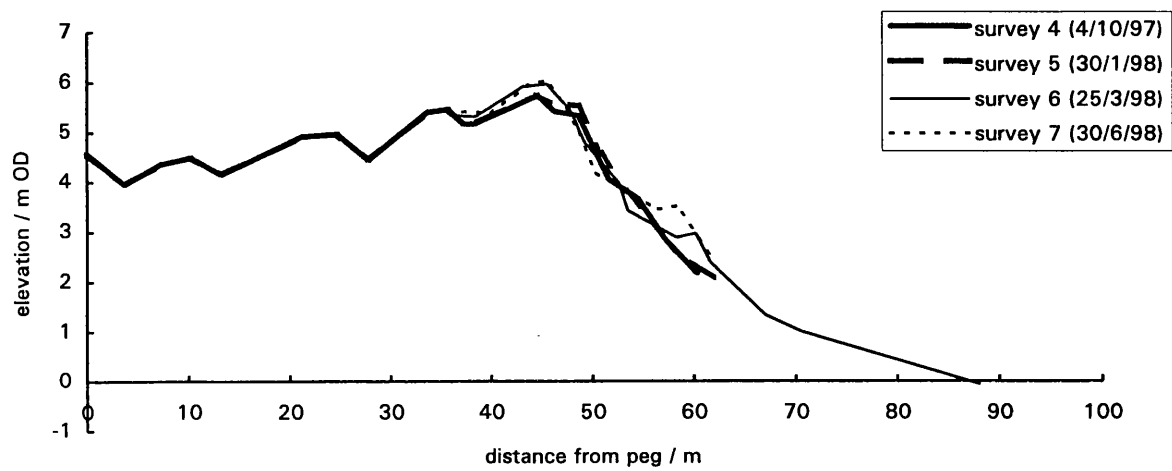
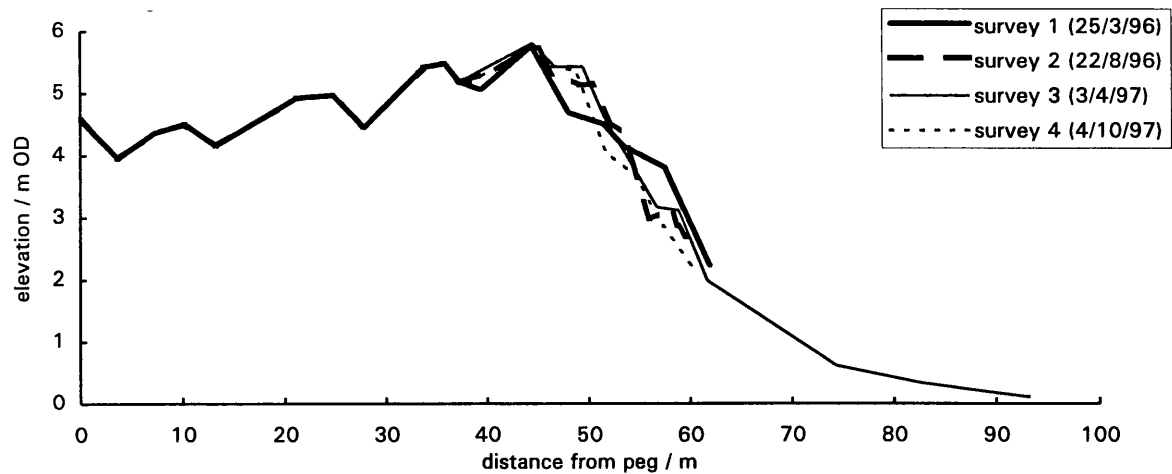
Profile -8: Morphological Change (March 1996 to March 1999)



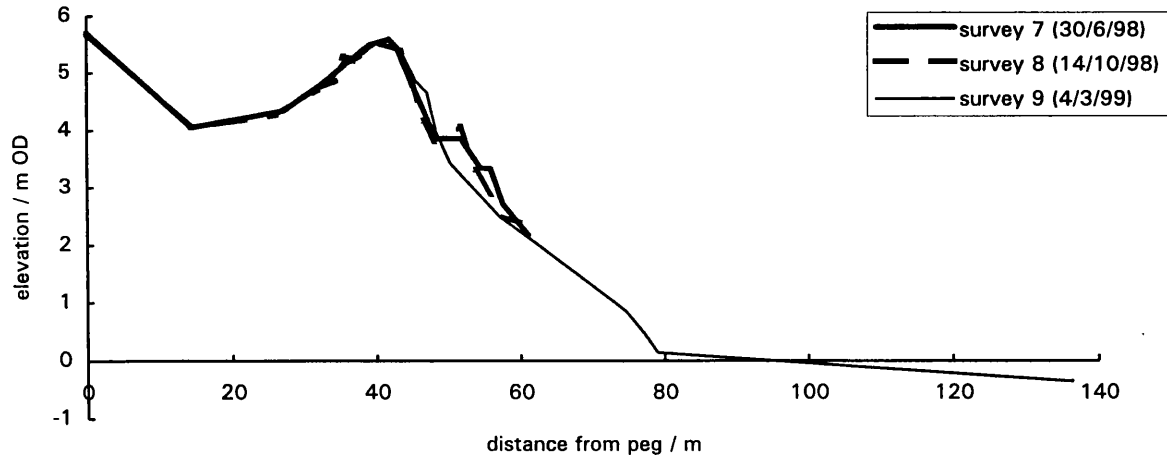
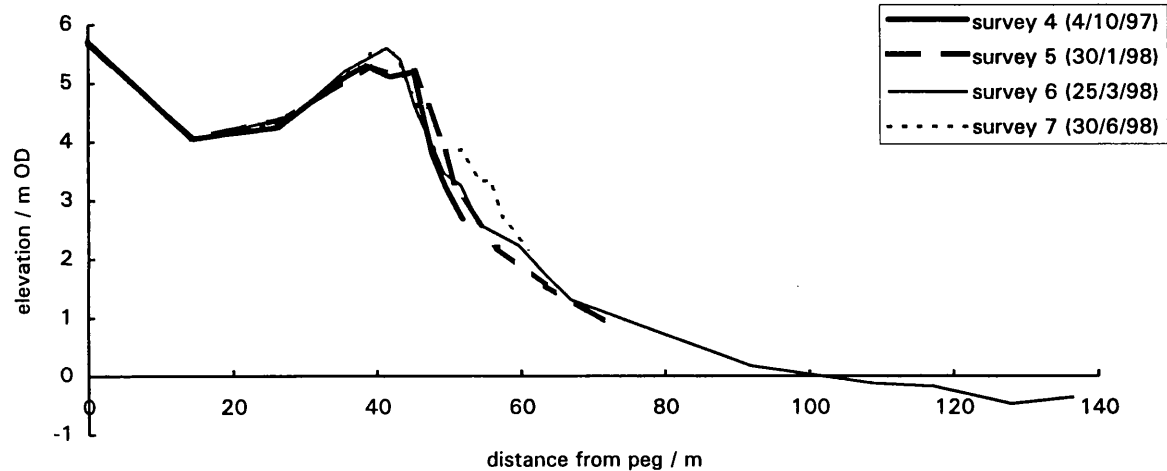
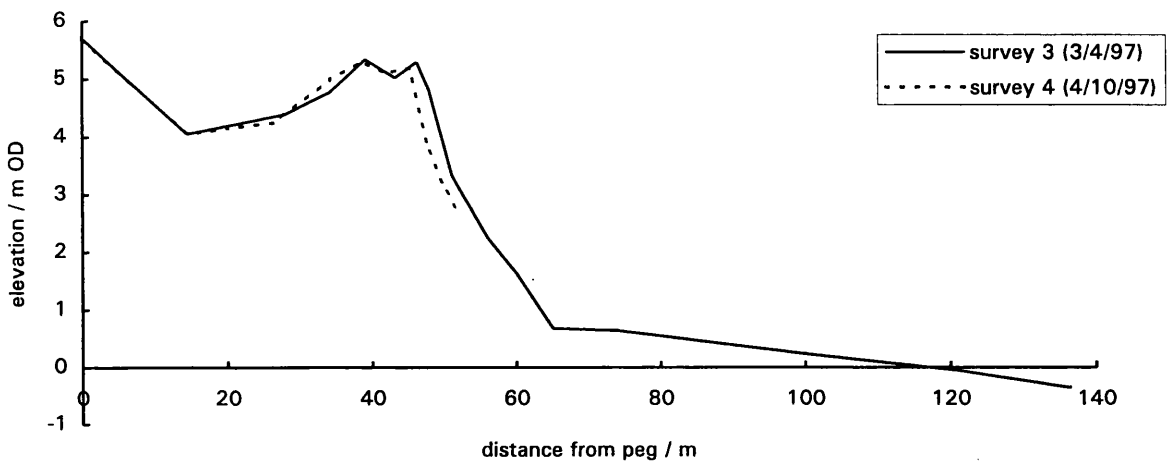
Profile -8.5: Morphological Change (March 1996 to March 1999)



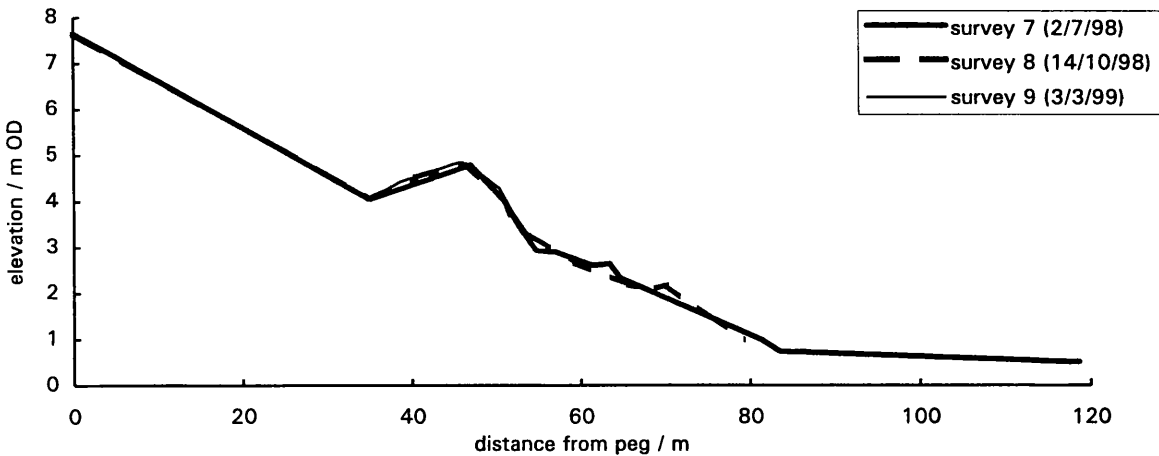
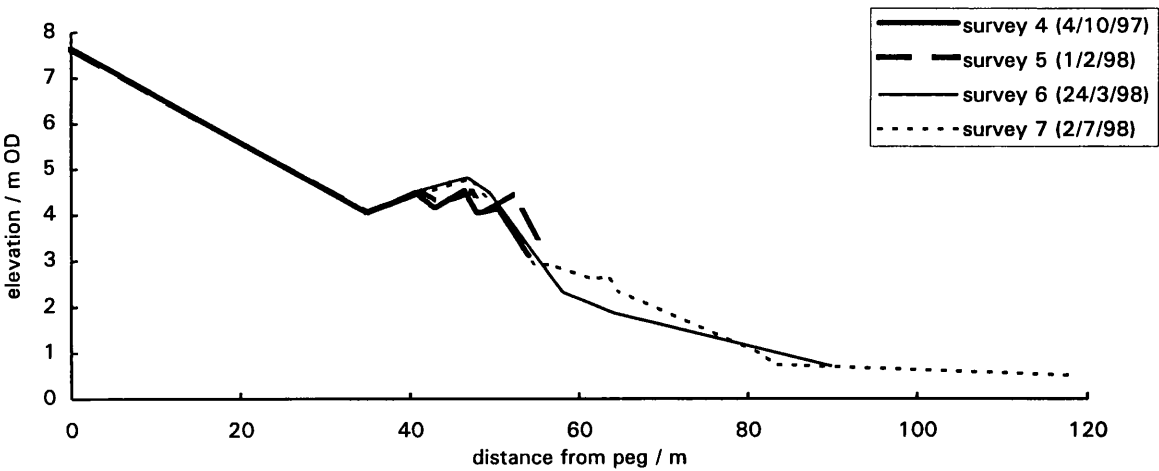
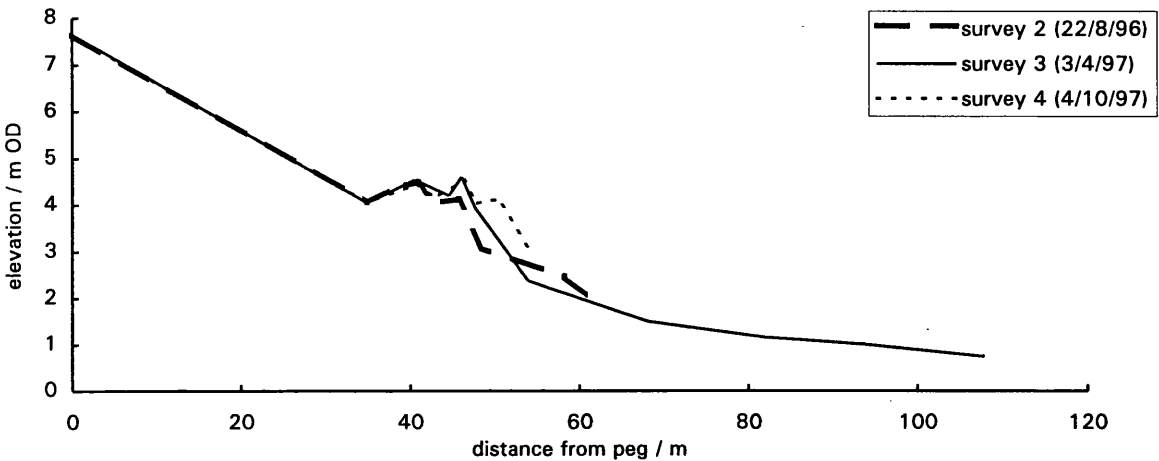
Profile -9: Morphological Change (March 1996 to March 1999)



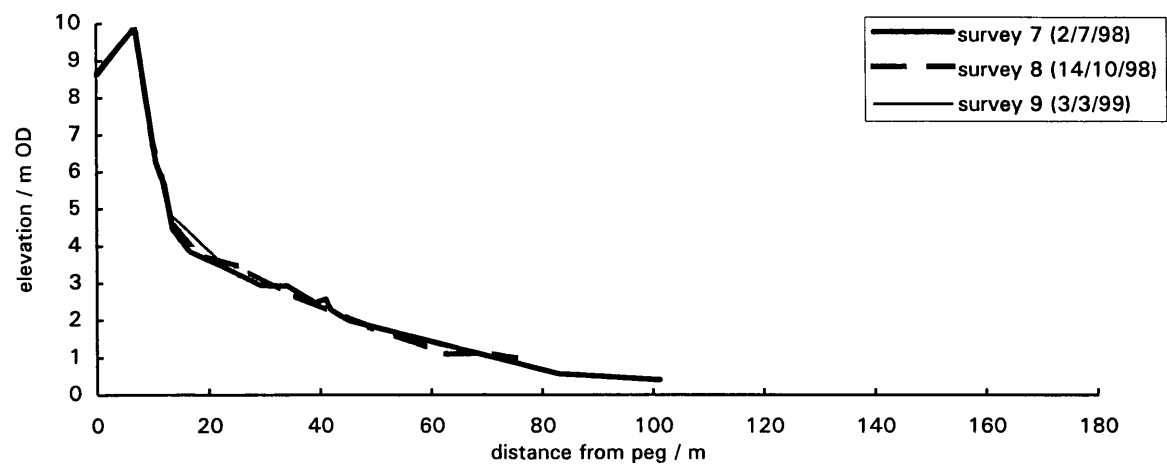
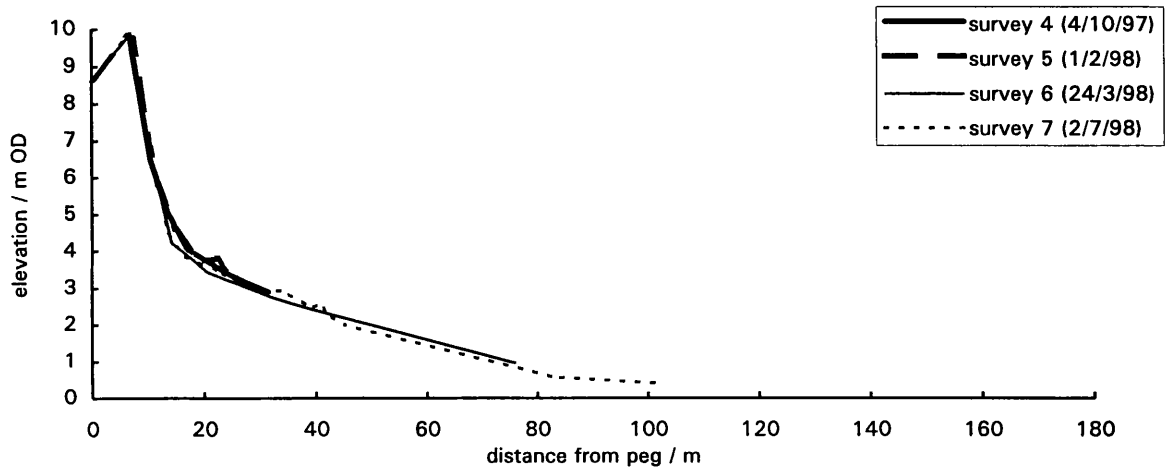
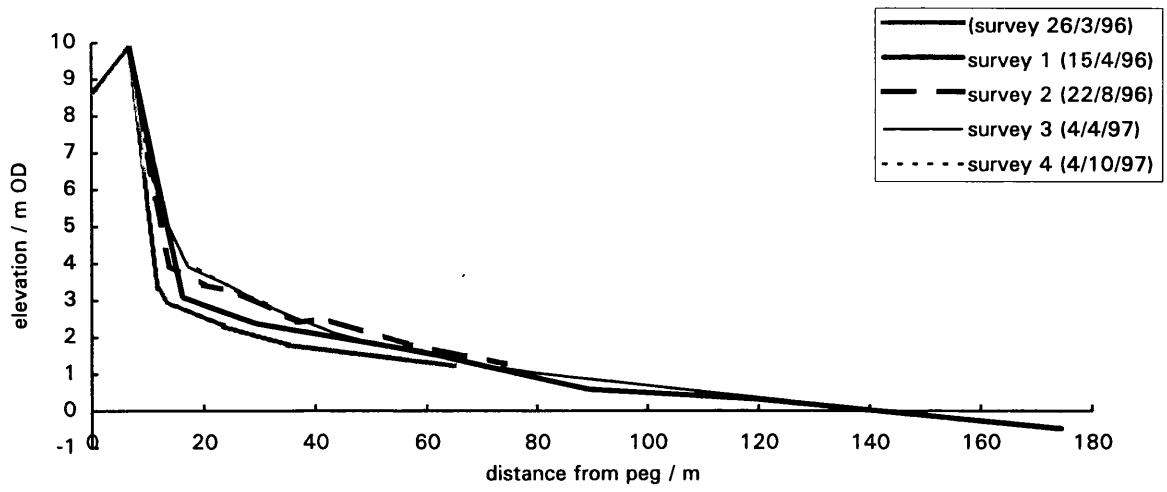
Profile -9.5: Morphological Change (April 1997 to March 1999)



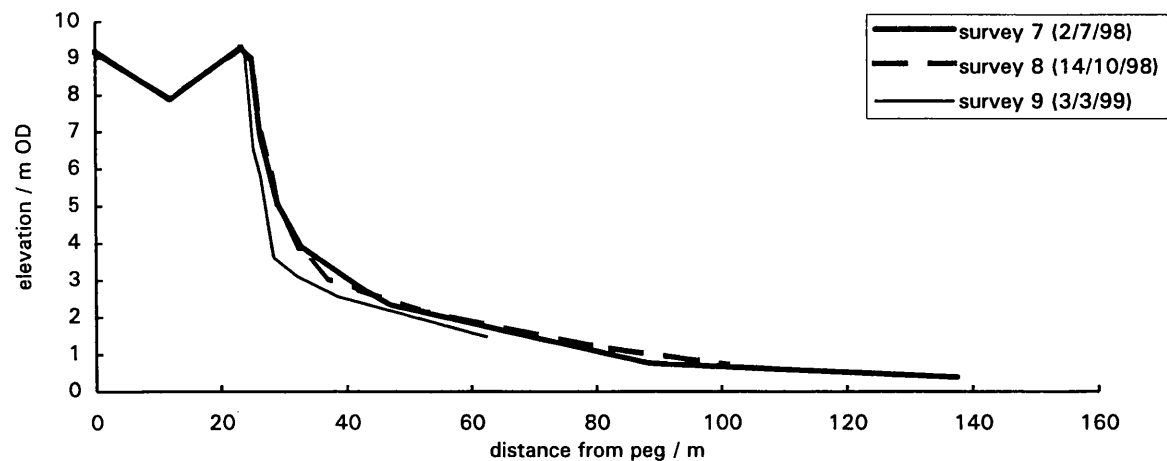
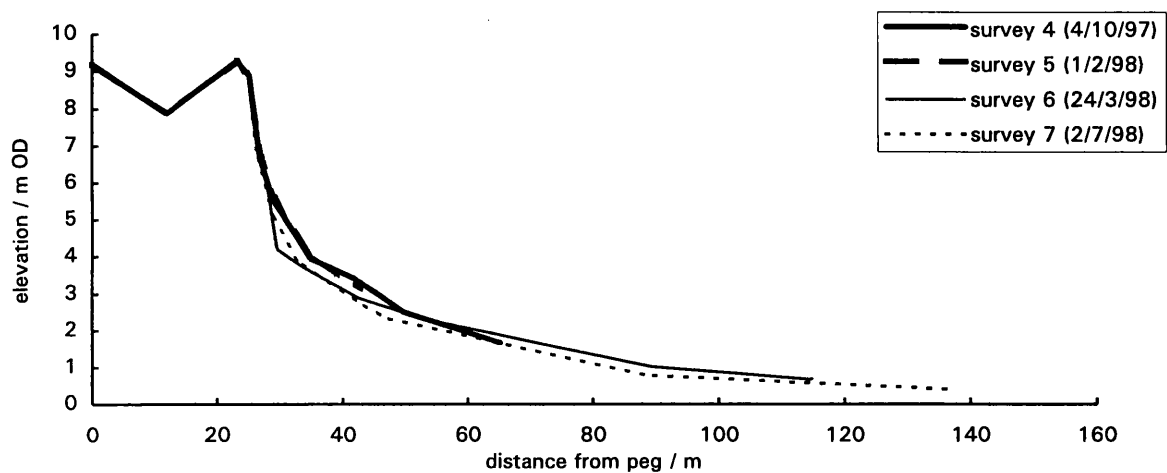
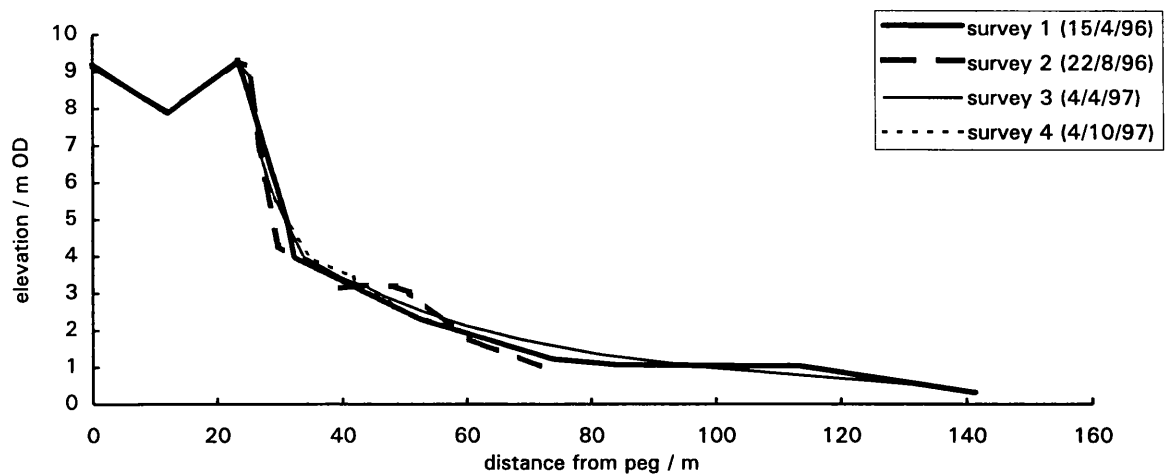
Profile -9.75: Morphological Change (August 1996 to March 1999)



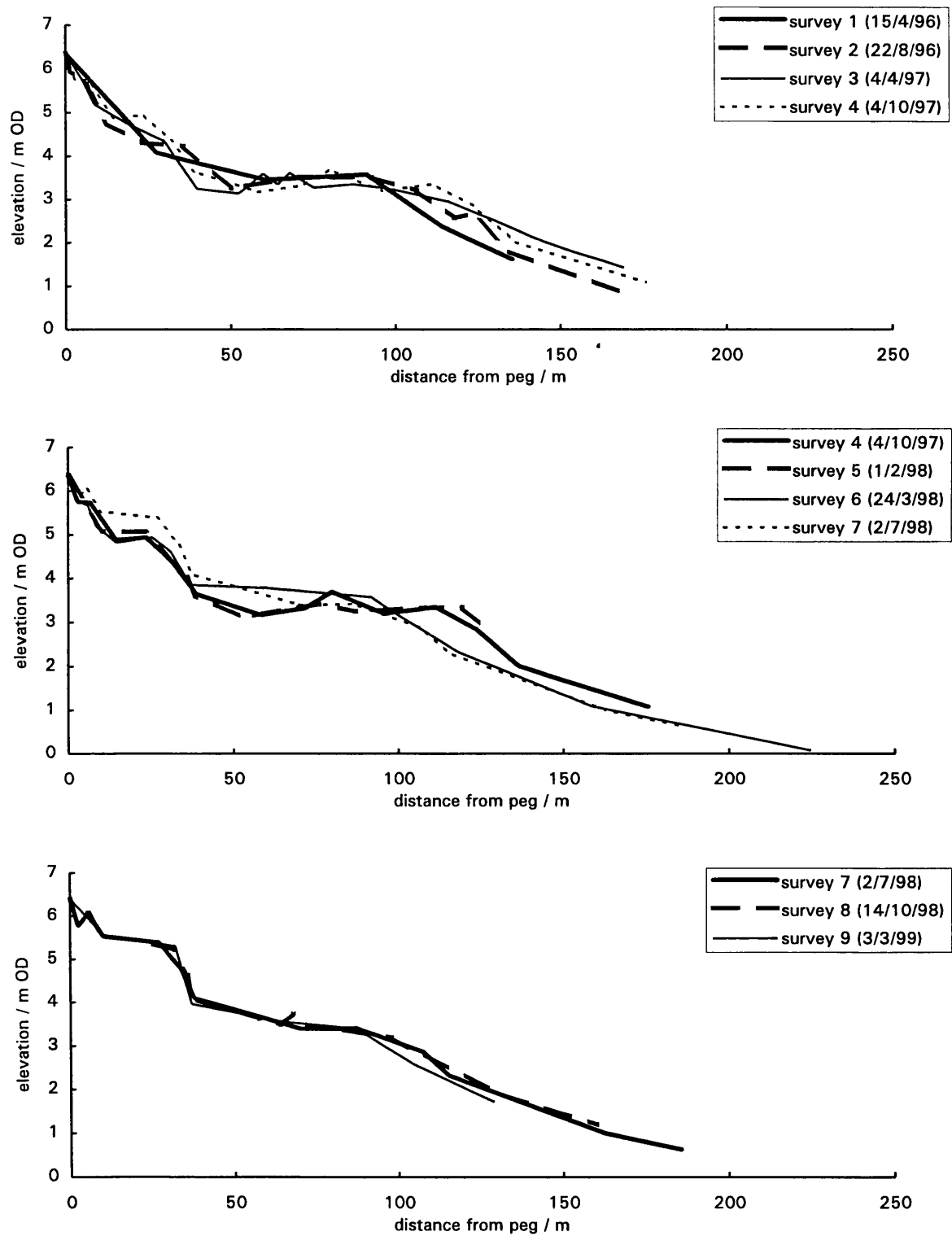
Profile -9.8: Morphological Change (March 1996 to March 1999)



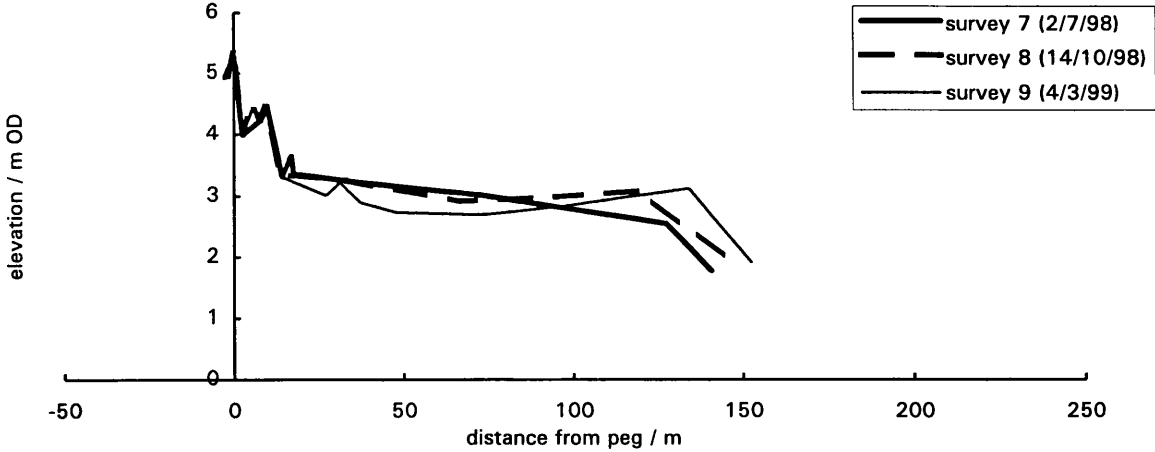
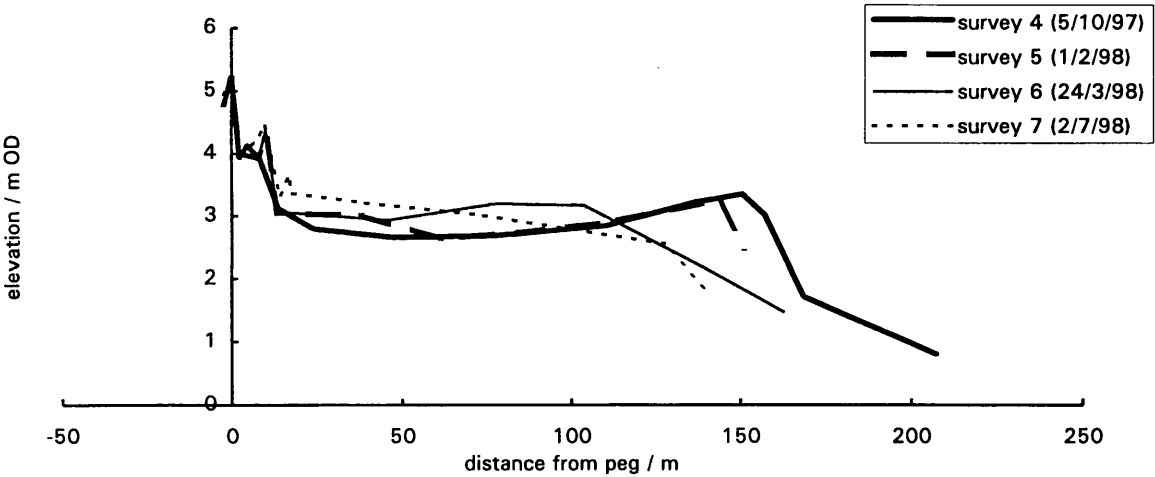
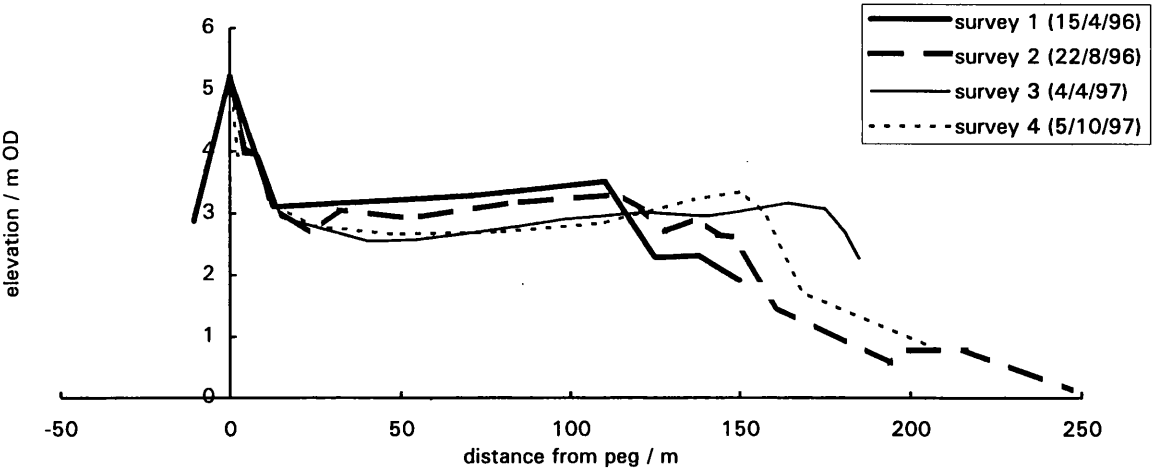
Profile -10.5: Morphological Change (April 1996 to March 1999)



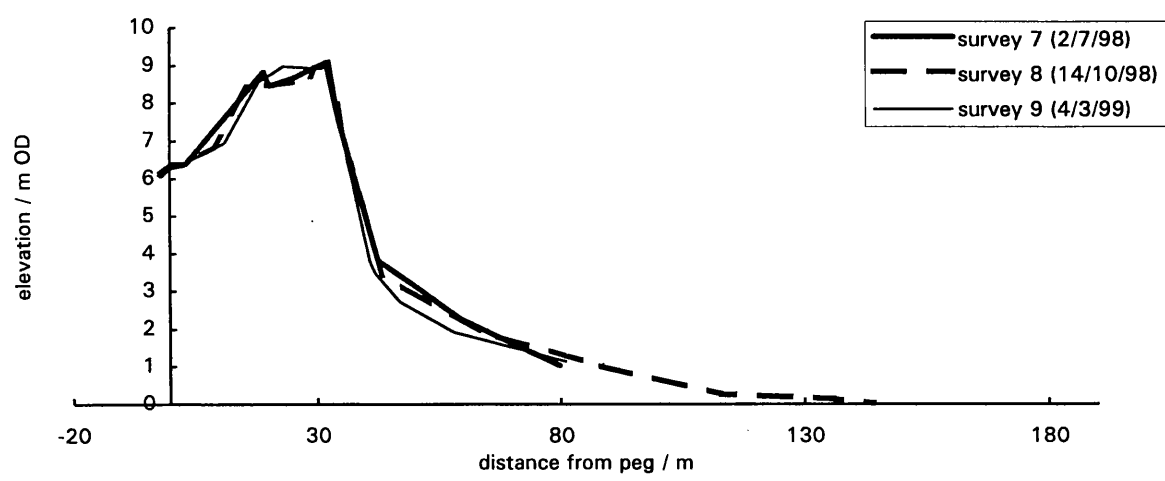
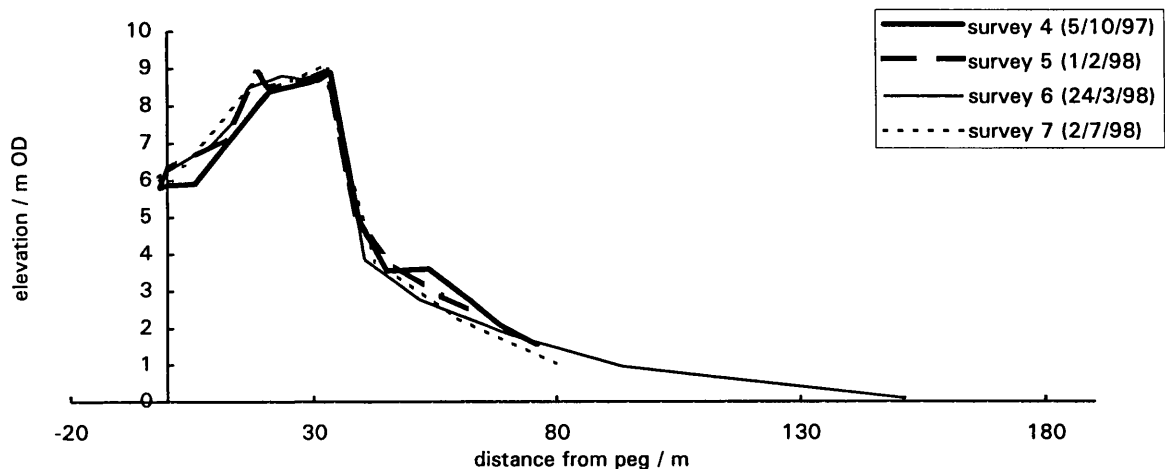
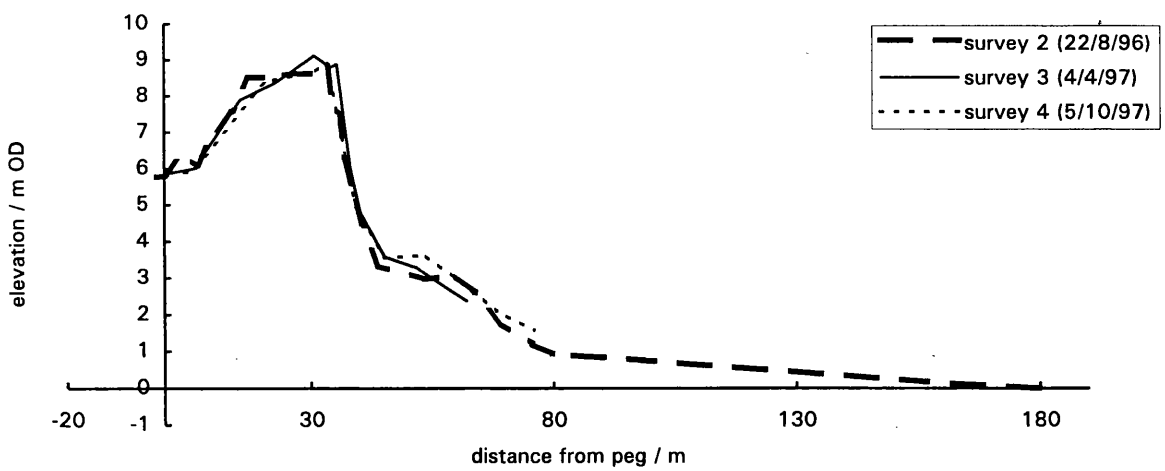
Profile -11: Morphological Change (April 1996 to March 1999)



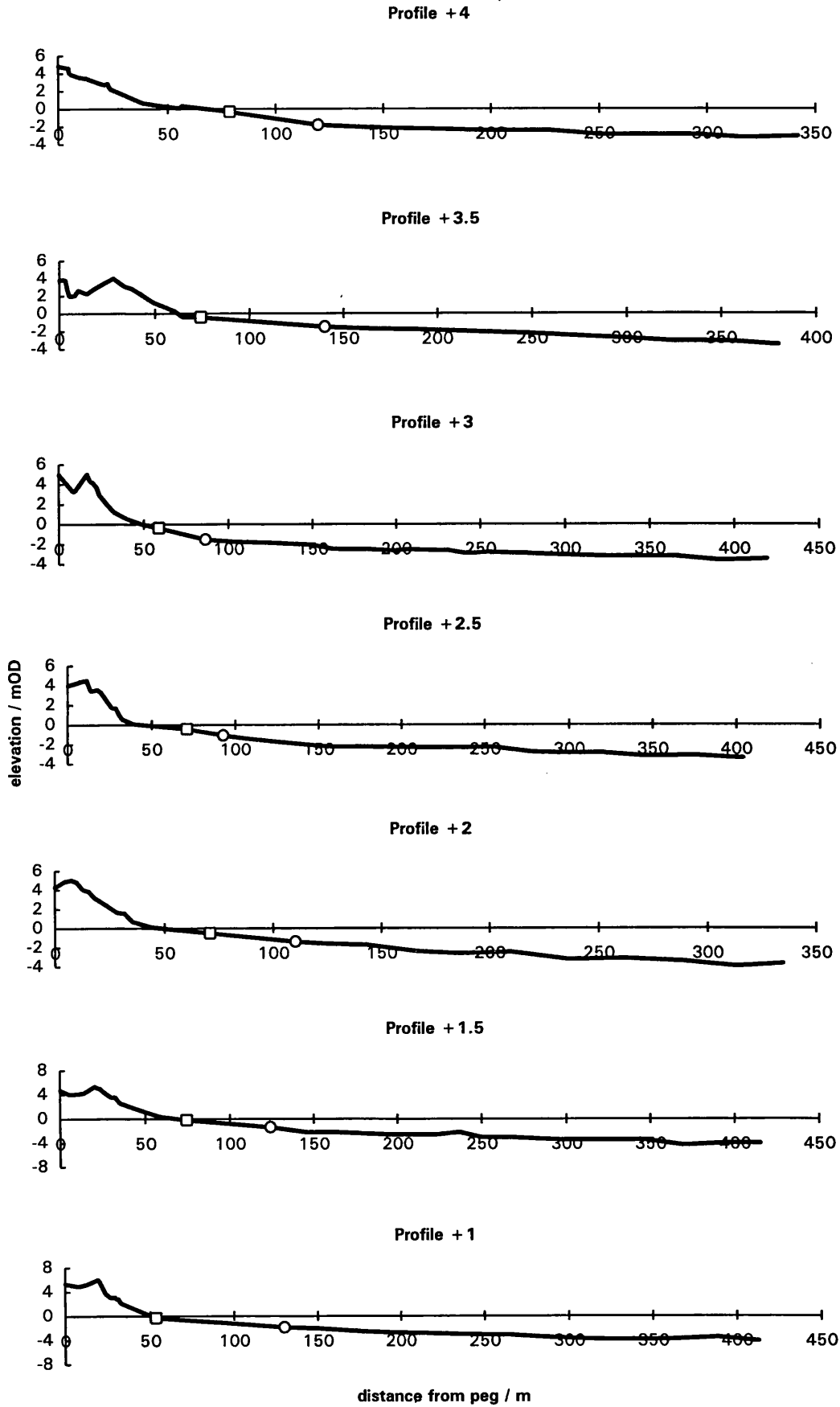
Profile -11.5: Morphological Change (April 1996 to March 1999)



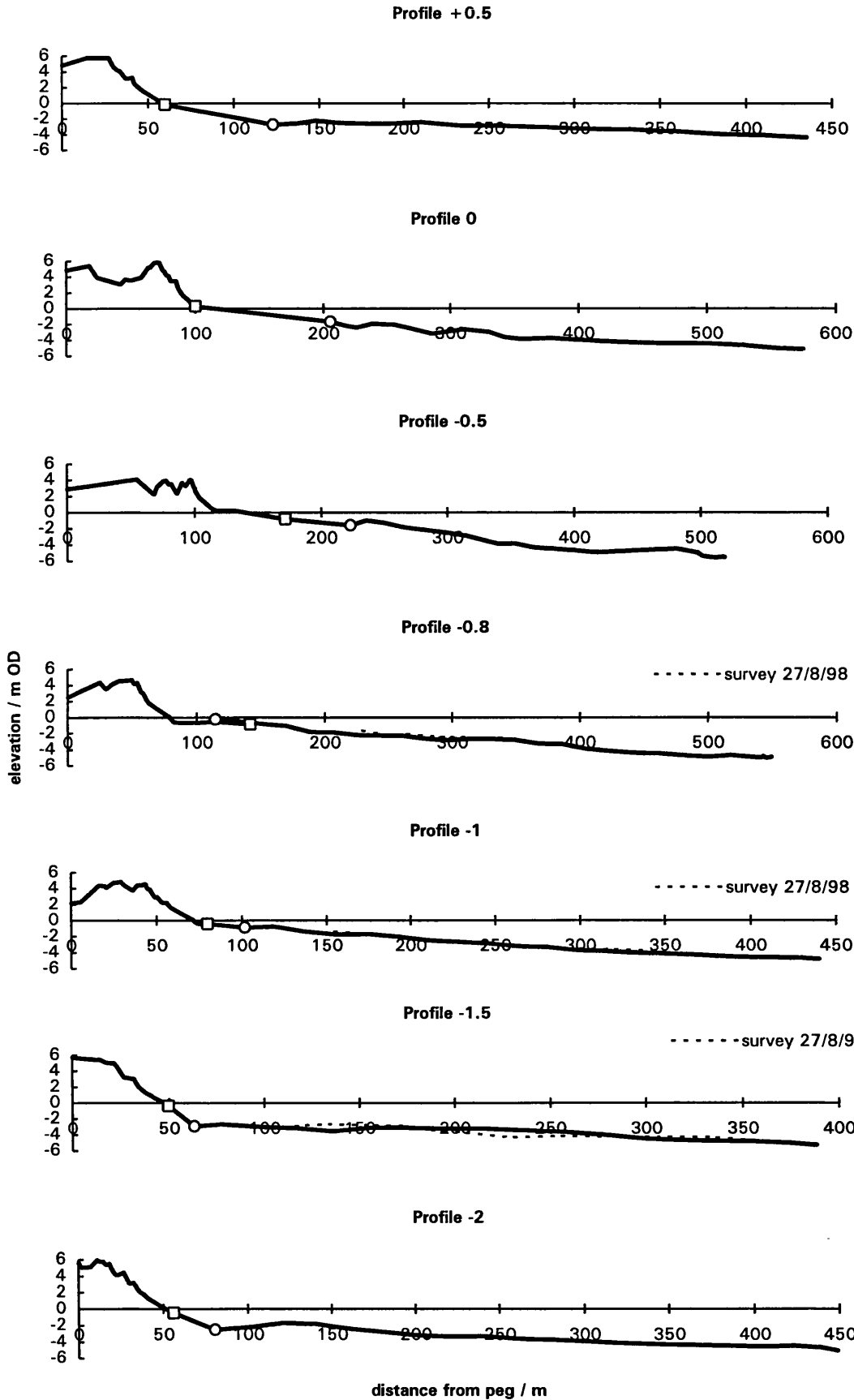
Profile -12: Morphological Change (April 1996 to March 1999)



APPENDIX G: Beach profiles extended offshore. Boxes and circles show the limits of the onshore and offshore surveys, respectively.

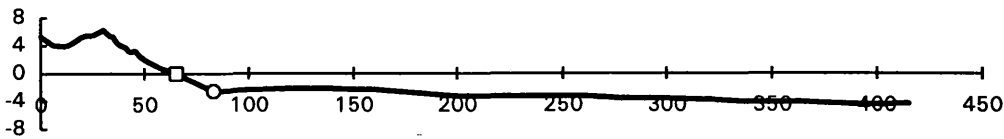


APPENDIX G (continued): Beach profiles extended offshore

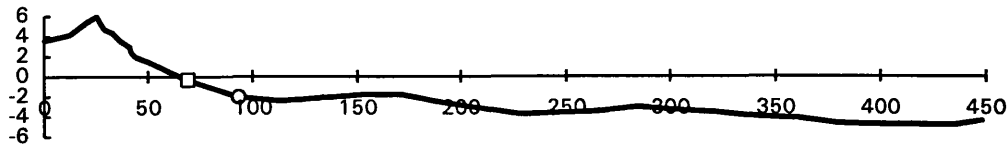


APPENDIX G (continued): Beach profiles extended offshore

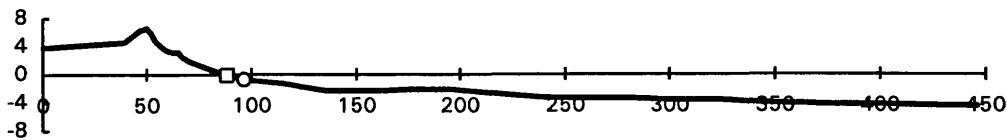
Profile -2.5



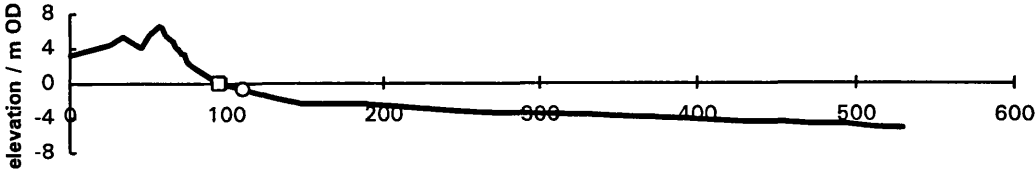
Profile -3



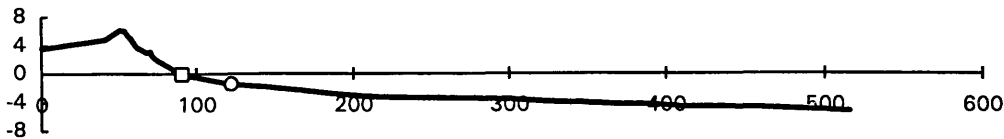
Profile -3.5



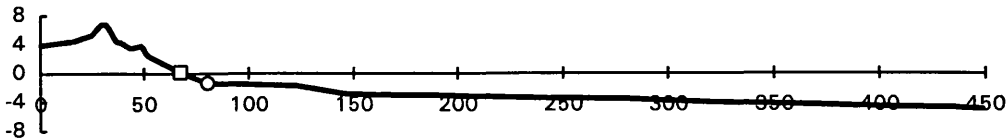
Profile -4



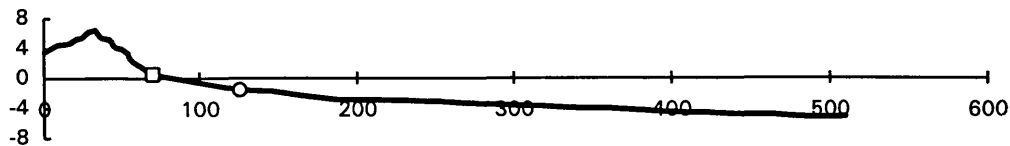
Profile -4.5



Profile -5



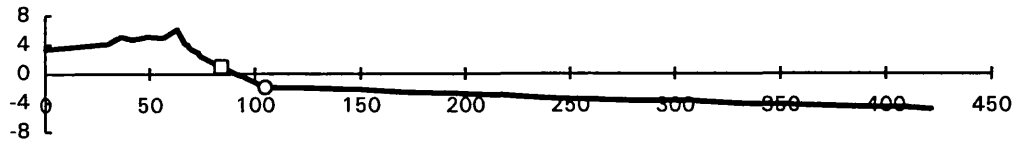
Profile -5.5



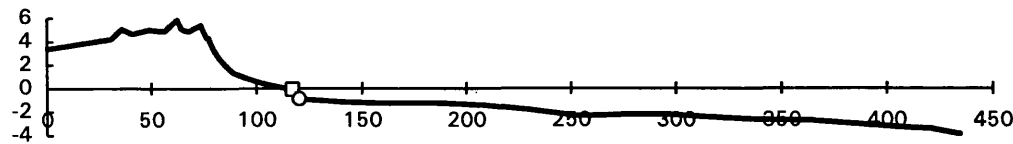
distance from peg / m

APPENDIX G (continued): Beach profiles extended offshore

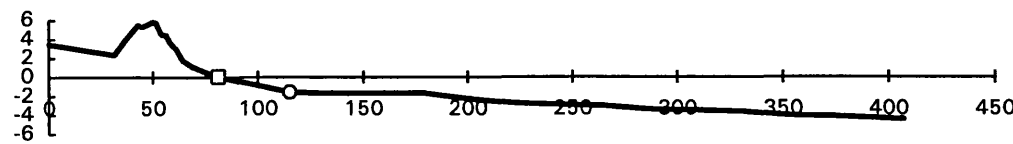
Profile -6



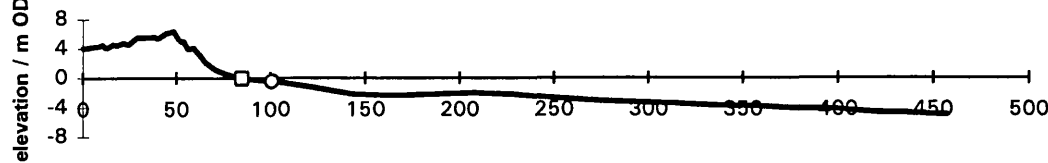
Profile -6.5



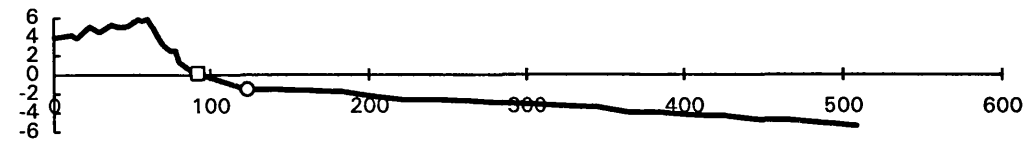
Profile -7



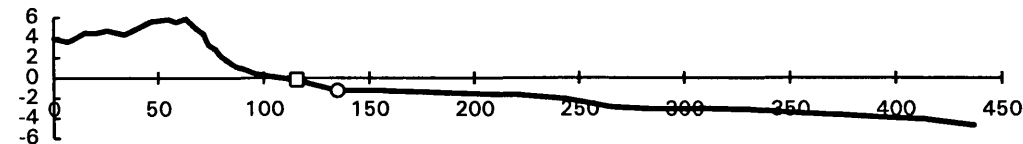
Profile -7.5



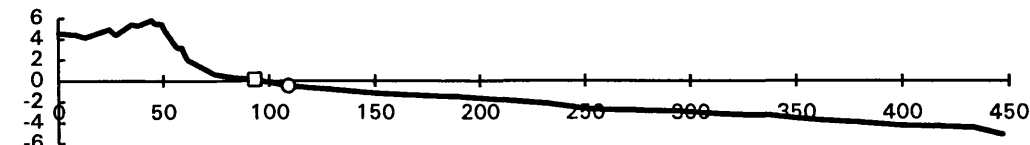
Profile -8



Profile -8.5



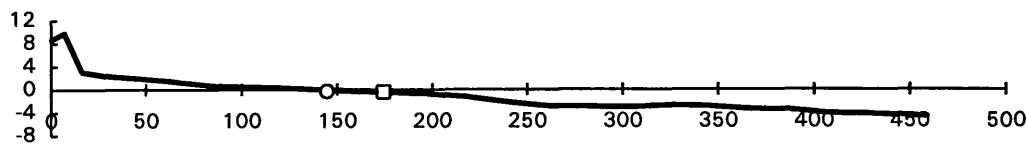
Profile -9



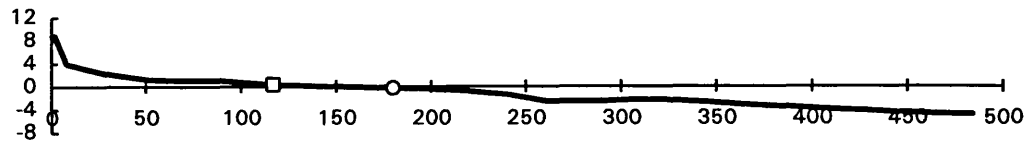
distance from peg / m

APPENDIX G (continued): Beach profiles extended offshore

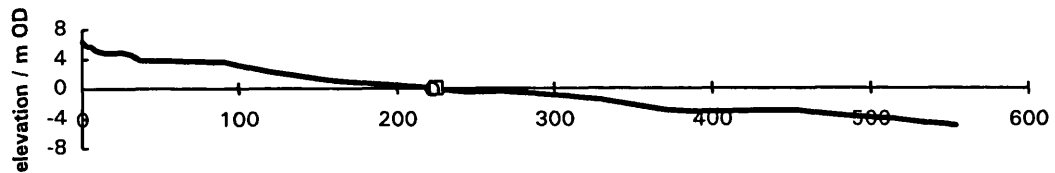
Profile -9.8



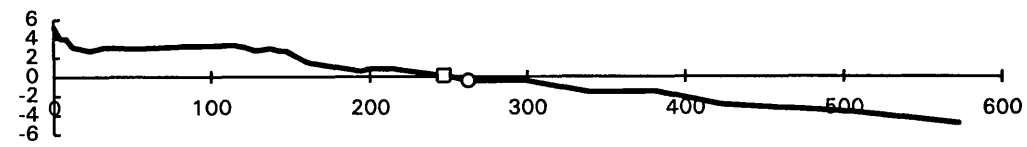
Profile -10.5



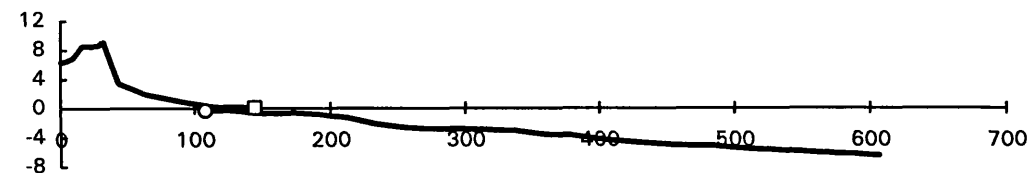
Profile -11



Profile -11.5



Profile -12



distance from peg / m

APPENDIX H: Summary Beach Surface Particle Size Analysis

Upper beach samples

profiles	mean (phi)	mean (mm)	sorting (phi)	skewness (phi)	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	$\sqrt{D84/D16}$ (mm)
-9	-5.35	40.79	0.65	-0.08	-5.39	41.93	-4.67	25.46	-6.00	64.00	1.59
-8	-5.18	36.25	0.63	0.05	-5.16	35.75	-4.56	23.59	-5.81	56.10	1.54
-7	-5.36	41.07	0.51	0	-5.37	41.36	-4.83	28.44	-5.88	58.89	1.44
-6	-5.17	36.00	0.57	-0.11	-5.20	36.76	-4.56	23.59	-5.76	54.19	1.52
-5	-5.09	34.06	0.44	0.13	-5.08	33.82	-4.65	25.11	-5.56	47.18	1.37
-4	-5.38	41.64	0.55	-0.09	-5.43	43.11	-4.78	27.47	-5.93	60.97	1.49
-3	-5.19	36.50	0.58	0.37	-5.05	33.13	-4.64	24.93	-5.89	59.30	1.54
-2	-5.28	38.85	0.60	0	-5.28	38.85	-4.67	25.46	-5.89	59.30	1.53
-1	-5.03	32.67	0.67	0.22	-4.93	30.48	-4.39	20.97	-5.77	54.57	1.61
0	-5.61	48.84	0.52	-0.09	-5.63	49.52	-5.09	34.06	-6.12	69.55	1.43
1	-5.04	32.90	0.62	-0.02	-5.04	32.90	-4.38	20.82	-5.71	52.35	1.59
2	-5.29	39.12	0.71	-0.16	-5.38	41.64	-4.50	22.63	-5.98	63.12	1.67
3	-5.17	36.00	0.64	0	-5.15	35.51	-4.54	23.26	-5.81	56.10	1.55
4	-5.27	38.59	0.74	0.05	-5.23	37.53	-4.52	22.94	-6.06	66.72	1.71
mean	-5.24	38.09	0.60	0.02	-5.24	38.02	-4.63	24.91	-5.87	58.74	1.54

Mid beach samples

profiles	mean (phi)	mean (mm)	sorting	skewness	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	$\sqrt{D84/D16}$ (mm)
-2.5	-5.54	46.53	0.51	0.22	-5.48	44.63	-5.09	34.06	-6.05	66.26	1.39
-2	-5.74	53.45	0.46	0	-5.74	53.45	-5.25	38.05	-6.24	75.58	1.41
-1.5	-5.10	34.30	0.64	0.27	-4.97	31.34	-4.53	23.10	-5.81	56.10	1.56
-1	-5.79	55.33	0.61	0.11	-5.75	53.82	-5.18	36.25	-6.45	87.43	1.55
-0.5	-5.17	36.00	0.60	-0.01	-5.16	35.75	-4.57	23.75	-5.77	54.57	1.52
1.5	-5.02	32.45	0.48	0.21	-4.97	31.34	-4.60	24.25	-5.50	45.25	1.37
2	-5.66	50.56	0.45	0.03	-5.66	50.56	-5.18	36.25	-6.13	70.03	1.39
3	-5.90	59.71	0.55	-0.02	-5.89	59.30	-5.36	41.07	-6.43	86.22	1.45
3.5	-4.98	31.56	0.56	0.13	-4.93	30.48	-4.44	21.71	-5.57	47.50	1.48
4	-5.2	36.76	0.53	0.09	-5.19	36.50	-4.68	25.63	-5.73	53.08	1.44
mean	-5.41	43.66	0.54	0.10	-5.37	42.72	-4.89	30.41	-5.97	64.20	1.46

APPENDIX H: Summary Beach Surface Particle Size Analysis (continued)

Lower beach samples												
profiles	mean (phi)	mean (mm)	sorting	skewness	D50 (phi)	D50 (mm)	D16 (phi)	D16 (mm)	D84 (phi)	D84 (mm)	√D84/D16 (mm)	
-2.5	-4.89	29.65	0.70	0.01	-4.88	29.45	-4.17	18.00	-5.62	49.18		1.65
-1.5	-4.68	25.63	0.60	-0.05	-4.70	25.99	-4.06	16.68	-5.29	39.12		1.53
-1	-5.05	33.13	1.00	-0.12	-5.13	35.02	-3.92	15.14	-6.11	69.07		2.14
-0.5	-5.81	56.10	0.85	0.03	-5.80	55.72	-5.02	32.45	-6.63	99.04		1.75
1.5	-4.14	17.63	0.46	-0.1	-4.18	18.13	-3.68	12.82	-4.56	23.59		1.36
2	-4.19	18.25	0.58	0.13	-4.15	17.75	-3.63	12.38	-4.80	27.86		1.50
3	-4.52	22.94	0.54	0.22	-4.42	21.41	-4.03	16.34	-5.10	34.30		1.45
3.5	-4.25	19.03	0.74	0.24	-4.17	18.00	-3.59	12.04	-5.00	32.00		1.63
mean	-4.69	27.80	0.68	0.05	-4.68	27.68	-4.01	16.98	-5.39	46.77		1.63

APPENDIX I: Volumetric change in each ca. 500m wide cell along the coast of Spey Bay between surveys (m³)

Note: All volumes are given as bulk volumes and volumes are calculated using profiles to their surveyed limits

Cells	Width (m)	Volumetric change between surveys (m ³)								Net change
		(1-2)	(2-3)	(3-4)	(4-5)	(5-6)	(6-7)	(7-8)	(8-9)	
-12-11.5	502	-1638	8701	-7124	3293	-2626	-130	1829	-5546	-3241
-11.5-11	508	-168	11983	-4775	1701	465	289	4151	-4720	8926
-11-10.5	520	1109	6417	1211	23	-1156	-2811	2193	-8108	-1122
-10.5-9.8	604	4378	2446	-818	731	-4409	-5230	996	-5985	-7890
-9.8-9.75	151	1216	57	218	405	-426	174	11	249	1905
-9.75-9.5	223	0	137	-36	743	356	1115	-154	-41	2121
-9.5-9	491	66	398	-2083	1215	1220	1823	-287	-2127	223
-9-8.5	380	-701	1635	-1054	695	173	-30	194	-1190	-279
-8.5-8	704	-746	3532	-2174	2051	-788	-1312	-580	232	215
-8-7.5	465	582	1284	-1412	1063	-209	-1961	44	-342	-951
-7.5-7	529	1408	4730	-4158	66	-6749	56	344	614	-3690
-7-6.5	512	2503	7183	-5910	-82	-5626	-2152	1547	958	-1578
-6.5-6	446	1047	3227	-5169	-451	1687	-1431	-1223	-710	-3022
-6-5.5	476	-395	2135	-3394	-1431	1076	1559	-3780	-1121	-5351
-5.5-5	367	325	1951	-275	-447	877	-895	-1636	-789	-889
-5-4.5	564	448	1643	-1674	211	1821	-753	-2469	-1291	-2063
-4.5-4	582	-425	715	-2184	-137	1335	849	-3527	-3088	-6461
-4-3.5	455	-1346	-300	-2508	1171	1496	-102	-3177	-2402	-7168
-3.5-3	583	-4134	-257	-2236	2376	2105	-971	-2230	-1163	-6510
-3-2.5	554	-2534	-119	-983	2332	1709	-1285	266	535	-80
-2.5-2	437	1112	306	-1171	1179	1424	-1044	22	745	2573
-2-1.5	588	3793	-1287	-2157	-782	1469	-2597	353	4252	3044
-1.5-1	376	470	-755	-2104	-365	-787	-3611	534	2934	-3683
-1-0.8	142	-519	496	-455	-572	-755	-971	-389	282	-2883
-0.5-0.8	361	-1420	4601	-1316	896	781	9789	-5212	-10768	-2649
0.1-0	74	-63	73	-130	880	694	-34	-64	-676	680
0-0.5	488	-200	2040	-2455	2560	3924	1540	-1599	-969	4842
0.5-1	437	157	4340	-811	1319	1621	-3561	-7	1105	4162
1-1.5	456	-1915	2893	-1503	371	1767	-6382	1562	1608	-1599
1.5-2	543	-1853	750	-1636	-160	2328	-5627	1657	2633	-1907
2-2.5	517	-1538	1277	-2782	349	1308	-1300	1088	814	-783
2.5-3	542	-2472	2564	-2551	960	274	-1158	1396	-1838	-2825
3-3.5	524	-432	2503	-8299	4833	1327	-3851	1236	1141	-1542
3.5-4	526	3908	612	-10622	5024	418	-2228	76	2092	-720

APPENDIX I (CONTINUED): Volumetric change in each ca. 500m wide cell along the coast of Spey Bay between surveys (m³)

Note: All volumes are given as bulk volumes and volumes are calculated using profiles extrapolated to a maximum depth

Cells	Width (m)	Volumetric change between surveys (m ³)								Net change	
		(1-2)	(2-3)	(3-4)	(4-5)	(5-6)	(6-7)	(7-8)	(8-9)		
-12-11.5	502	-4946	34249	-22267	15048	-28447	-780	-208	2907	-4445	
-11.5-11	508	-5265	25484	-13996	13555	-24278	-1495	8784	-4080	-1292	
-11-10.5	520	-5217	15396	-255	19758	-28690	-3237	3715	-8449	-6978	
-10.5-9.8	604	4484	4949	18561	13135	-18757	-6676	5139	19295	40130	
-9.8-9.75	151	2560	-1404	8011	1206	-4350	-145	1277	9349	16506	
-9.75-9.5	223	0	-985	10765	-2933	-6998	6444	900	-2210	4983	
-9.5-9	491	1087	-4272	15049	-9106	-4949	16006	-81	-16607	-2873	
-9-8.5	380	-3208	-2151	3860	3852	-7330	6276	-1934	-2217	-2853	
-8.5-8	704	-6902	4070	2902	12352	-14719	7472	-2850	-1302	1024	
-8-7.5	465	1458	100	-1091	6536	-4289	-2296	337	-312	444	
-7.5-7	529	-813	2197	-3994	3991	-9970	-985	1294	2639	-5640	
-7-6.5	512	679	3935	-3630	8616	-14667	-9246	5195	6730	-2388	
-6.5-6	446	1611	-531	-2277	7176	-6165	-7122	1091	978	-5238	
-6-5.5	476	-771	-38	-1419	-1973	-888	2378	-4725	-2830	-10266	
-5.5-5	367	1401	642	1984	-2850	836	-152	-2193	-1639	-1971	
-5-4.5	564	2419	400	1629	-2630	4011	-1643	-3002	-2187	-1003	
-4.5-4	582	869	-756	-2413	384	3757	-2617	-4015	-3338	-8130	
-4-3.5	455	-1517	-755	-5752	6359	2622	-3195	-3727	-2609	-8574	
-3.5-3	583	-6113	8655	-9396	8133	6367	-3968	-4548	-755	-1624	
-3-2.5	554	-3825	7980	-9268	7152	4379	-4398	-1310	1446	2155	
-2.5-2	437	1103	3226	-8227	5174	1909	-4333	566	1699	1116	
-2-1.5	588	3820	3370	-9453	435	3649	-8750	3458	6764	3294	
-1.5-1	376	-704	4298	-8013	-561	-657	-7665	2087	4268	-6946	
-1-0.8	142	-4819	5664	-5872	-916	-681	-1211	-1454	-3161	-12451	
-0.5-0.8	361	-11784	11196	-16339	6243	1287	19543	-9733	-21057	-20644	
0.1-0	74	-15	-3531	-896	1305	208	-19	-200	-1035	-4183	
0-0.5	488	-683	-6939	-3990	3927	5	2148	-1630	-2802	-9965	
0.5-1	437	-483	5339	1691	3626	-816	-3220	1301	1966	9405	
1-1.5	456	-863	1795	1237	7377	-6671	-6195	5512	4410	6601	
1.5-2	543	2728	-7909	6727	15117	-20849	-5681	10974	6083	7189	
2-2.5	517	6585	-8547	6140	15921	-26490	-1502	16125	2816	11047	
2.5-3	542	3820	-2814	2789	11304	-16452	-2001	8371	1857	6875	
3-3.5	524	424	592	-6636	8152	-801	-3838	-1869	4788	812	
3.5-4	526	3724	912	-12117	5576	184	-1387	-173	2888	-394	

APPENDIX J : Maximum error (in m³) in beach cell volume change calculations

Note: errors given as bulk volumes

Cells	Width (m)	Cell Volume Change Error (m ³)									
		(1-2)	(2-3)	(3-4)	(4-5)	(5-6)	(6-7)	(7-8)	(8-9)		
-12-11.5	502	1934	10269	8408	3888	3101	176	2162	6546		
-11.5-11	508	235	14142	5638	2008	555	356	4900	5571		
-11-10.5	520	1313	7574	1432	30	1370	3321	2590	9570		
-10.5-9.8	604	5169	2888	968	864	5205	6173	1179	7064		
-9.8-9.75	151	1435	69	258	479	503	208	19	295		
-9.75-9.5	223	0	162	61	878	421	1317	183	56		
-9.5-9	491	79	470	2460	1435	1441	2153	345	2512		
-9-8.5	380	828	1930	1245	822	211	66	230	1406		
-8.5-8	704	888	4171	2568	2423	933	1551	687	278		
-8-7.5	465	688	1517	1668	1256	249	2316	80	405		
-7.5-7	529	1663	5583	4908	95	7965	155	410	729		
-7-6.5	512	2956	8479	6976	106	6641	2545	1828	1132		
-6.5-6	446	1237	3810	6101	535	1993	1695	1450	839		
-6-5.5	476	468	2521	4007	1690	1271	1843	4462	1325		
-5.5-5	367	387	2304	327	531	1036	1058	1932	932		
-5-4.5	564	535	1942	1978	253	2151	890	2916	1526		
-4.5-4	582	503	846	2579	164	1577	1004	4164	3646		
-4-3.5	455	1590	356	2962	1384	1768	148	3751	2836		
-3.5-3	583	4880	308	2640	2806	2486	1148	2634	1375		
-3-2.5	554	2992	149	1162	2754	2019	1518	326	639		
-2.5-2	437	1314	365	1383	1393	1682	1233	80	881		
-2-1.5	588	4479	1524	2547	924	1736	3067	431	5019		
-1.5-1	376	562	894	2484	433	931	4263	634	3463		
-1-0.8	142	613	586	538	675	892	1146	460	334		
-0.5-0.8	361	1678	5430	1553	1064	926	11552	6151	12708		
0.1-0	74	76	87	154	1038	819	46	76	798		
0-0.5	488	255	2409	2899	3023	4633	1821	1888	1148		
0.5-1	437	202	5123	959	1558	1914	4203	86	1305		
1-1.5	456	2262	3415	1776	441	2088	7533	1846	1899		
1.5-2	543	2189	887	1932	191	2750	6642	1958	3109		
2-2.5	517	1817	1509	3285	413	1546	1537	1287	966		
2.5-3	542	2920	3028	3012	1135	328	1369	1650	2171		
3-3.5	524	519	2955	9794	5704	1568	4546	1460	1354		
3.5-4	526	4613	727	12536	5931	501	2632	100	2471		

FIGURE 3.14: RIVER SPEY - GEOMORPHOLOGY

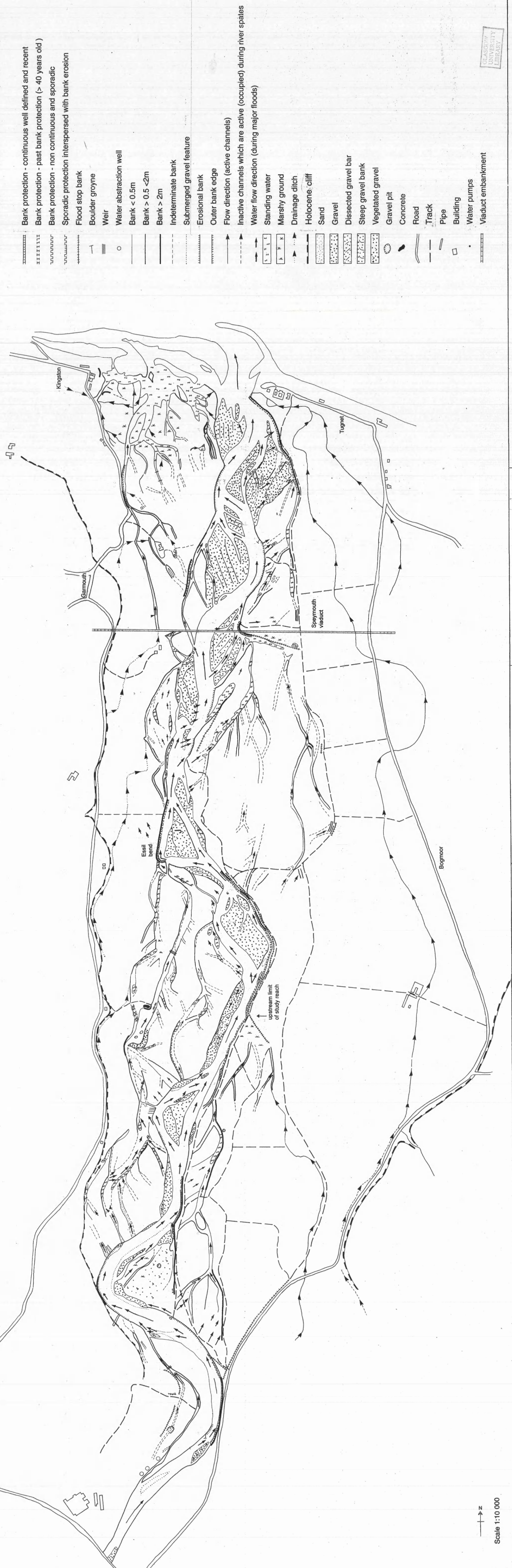


FIGURE 3.21: SPEY BAY - Geomorphology

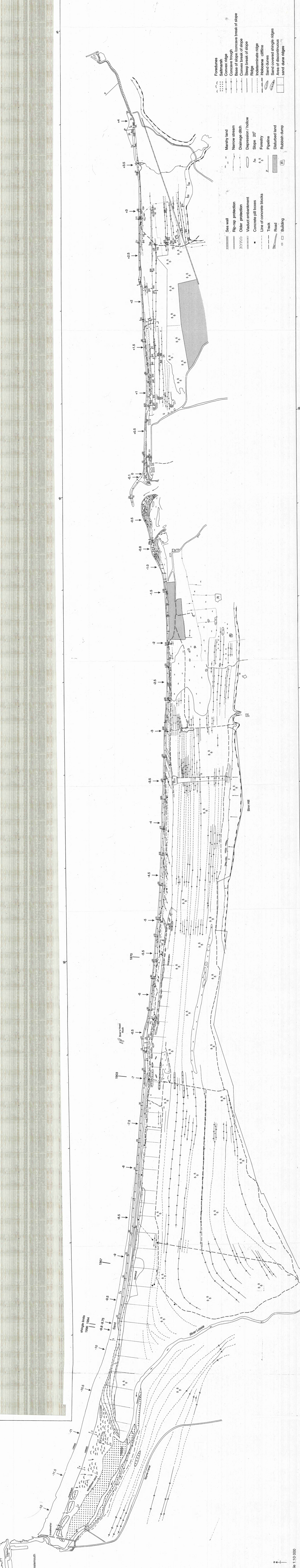
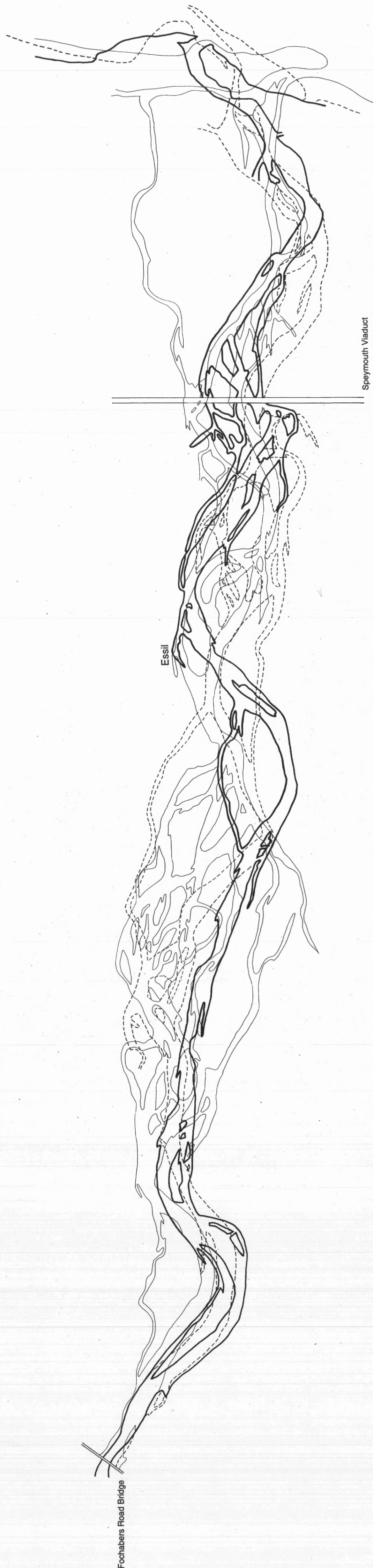


FIGURE 5.2: RIVER SPEY - PLANIMETRIC CHANNEL CHANGE, 1870, 1903, 1971



1870
1903
1971



Scale 1:10 560

FIGURE 6.1 : SPEY BAY SHORELINE POSITIONS (1870 AND 1970)

